THE UNIVERSITY OF ALBERTA

RELEASE FORM

NAME OF AUTHOR . . . BRIAN WILLIAM GLOVER . . .

TITLE OF THESIS . . . LATE QUATERNARY RIVER TERRACE EVOLUTION . . .

IN PART OF THE ATHABASCA RIVER VALLEY . . .

DEGREE FOR WHICH THESIS WAS PRESENTED . . . M. Sc. . . .

YEAR THIS DEGREE GRANTED . . . 1979 . . .

THE UNIVERSITY OF ALBERTA

LATE QUATERNARY RIVER TERRACE EVOLUTION IN PART OF THE ATHABASCA RIVER VALLEY

by

© BRIAN WILLIAM GLOVER

A THESIS

SUBMITTED TO THE FACULTY OF GRADUATE STUDIES AND RESEARCH

IN PARTIAL FULFILLMENT OF THE REQUIREMENTS FOR THE DEGREE

MASTER OF SCIENCE

DEPARTMENT OF GEOGRAPHY

EDMONTON, ALBERTA

FALL, 1979

UNIVERSITY OF ALBERTA

FACULTY OF GRADUATE STUDIES AND RESEARCH

The undersigned certify that they have read, and recommend to the Faculty of Graduate Studies and Research for acceptance, a thesis entitled "Late Quaternary River Terrace Evolution in Part of the Athabasca River Valley," submitted by Brian William Glover in partial fulfillment of the requirements for the degree of Master of Science.

ABSTRACT

This study primarily encompasses the mapping of remnant terrace distributions in order to establish their relative positions along the Athabasca River between Hinton and Whitecourt. The probable relationship of these features to major, proximal and distal, glacial controlling factors is considered. Terrace elevations above the present river channel and their distribution through the valley were determined by altimetric surveying and aerial photograph analysis. Terrace alluvium is described and predominant grain sizes were determined by a photographic grid-by-number technique on embedded grain surfaces.

Four paired terrace surfaces were identified within this section of the river, the oldest and uppermost terrace surface terminating approximately 20 kilometers downstream of Hinton. The remaining three terraces each of much greater downvalley extent, are divergent in nature. The relative gradients of these three terrace surfaces extend downvalley to the approximate elevations of three remnant Laurentide proglacial lake delta's, earlier identified by St-Onge (1972). The elevation, continuity and grain size distributions of the four terrace sets are considered to have been controlled by Cordilleran glacier sediment/discharge relationships in the upstream valley sector and/or episodic, Laurentide, proglacial

lake development downstream. Combined with limiting radiocarbon dates, and pertinent related studies, the alluvial terrace data are used to develop an interpretation of the apparently simultaneous sequence of glacial recession in this area during the late Quaternary.

ACKNOWLEDGMENTS

So many people were instrumental to the completion of this study, to thank them all here would be next to impossible. However, several people do deserve special mention and my warmest thanks.

First, and most importantly, I would like to thank Dr. R.B. (Bruce) Rains who so patiently supervised the research and writing of this study. His keen interest and understanding in my attempt to unravel the very complex Quaternary history of one region of Alberta helped me to complete this study and encouraged me to come back to try and learn more. More importantly though, his continuous enthusiasm for my work and friendship during those periods when quitting seemed to be the best alternative, will always be greatly appreciated.

To Debbie Pike who survived what many would consider to be rather unbearable field conditions and my cooking, a very special thanks. Her interest in my work and the very significant contributions she made to it both in the field, and later during the writing and compilation of this research helped dramatically in achieving its completion.

Appreciation is extended to Drs. J. Shaw and N. Rutter for reviewing this work and offering comments and suggestions on various sections. I'd also like to thank

Randi Pakan and Jack Chesterman for their technical advice and the many hours they put into the production of maps and photographs used in this study.

To the boys, Peter Ashmore, Olaf Niemann, Steve Luck, Dave Wright and Ken Thompson, my thanks for providing the pleasant, quite and enjoyable working conditions in the office. The sanity breaks were needed on more than one occasion.

Financial assistance for this research was provided by the Boreal Research Institute for Northern Studies and the University of Alberta Department of Geography and is gratefully acknowledged.

TABLE OF CONTENTS

Page

ABSTRACT IV

ACKNOWLEDGMENTS VI

LIST OF TABLES XI

LIST OF PLATES XV

INTRODUCTION 1

CHAPTER

I ALLUVIAL RIVER TERRACES 3

1.1 Causative Factors of Alluvial Terrace Development 3

1.1.1 Introduction 3

1.1.2 Climatic Controls 4

1.1.3 Tectonic/Isostatic Controls 12

1.1.4 Base Level Controls 17

1.2 Sedimentary Characteristics of Alluvial Terraces 21

1.2.1 Introduction 21

1.2.2 Meandering River Stratigraphy 21

1.2.3 Braided River Stratigraphy 23

1.2.4 General Conclusions 31

1.3 Alluvial Terraces in Alberta; Their Quaternary Significance 37

1.3.1 Introduction 37

CHAPTER Page

1.3.2 Alluvial Terrace Studies in Alberta 39

1.3.3 General Conclusions 58

II RESEARCH METHODS, HYDROLOGY AND BEDROCK GEOLOGY 61

2.1 Introduction 61

2.2 Research Methods 67

2.2.1 Problem Formulation 67

2.2.2 Field Methods 70

2.3 Hydrology 74

2.4 Bedrock and Structure 79

III SURFICIAL GEOLOGY 86

3.1 Introduction 86

3.2 General Glacial History 101

3.3 Major Surficial Deposit Units 109

3.3.1 Tills 109

3.3.2 Other Stratigraphic Units 116

3.4 General Conclusions 126

IV RIVER TERRACE AND DELTA CHARACTERISTICS 129

4.1 Introduction 129

4.2 Methodology 130

4.3 Terrace Characteristics 137

4.3.1 Morphology and Generalized Stratigraphy 137

4.3.2 Grain Size Analysis 167

CHAPTER		Page
	4.4 Delta Characteristics	171
V	INTERPRETATION AND CONCLUSIONS	178
	5.1 Introduction	178
	5.2 Interpretation	181
	5.3 Conclusions	190
BIBLIOGRAPHY	. . . .	199
APPENDIX A:	STRATIGRAPHIC SECTIONS	205
B:	FIELD AND DATA ANALYSIS PROCEDURES USED TO OBTAIN THE PREDICTED MEDIAN AND MEAN SIEVE DIAMETERS d_{sp50} & $\bar{d}_{sp50}$ . .	232
C:	EXAMPLES OF GRAIN SIZE CALCULATIONS	236
D:	FORMULAS APPLIED IN HOMOGENEITY OF VARIANCE AND DIFFERENCE OF MEANS TESTS . .	238

LIST OF TABLES

TABLE	DESCRIPTION	Page
1-1	EFFECTS OF TEMPERATURE AND PRECIPITATION ON SEDIMENT CONCENTRATION	10
1-2	PRINCIPLE DEPOSITIONAL FACIES OF BRAIDED STREAMS	27
1-3	PRINCIPLE DEPOSITIONAL PROCESSES OF BRAIDED STREAMS	29
1-4	WITHIN-CHANNEL BEHAVIOR OF MEANDERING AND BRAIDED STREAMS	32
1-5	TERRACES NORTH OF THE BOW RIVER, COCHRANE, ALBERTA	43
2-1	HYDROLOGICAL DATA, 1961 to 1976	76
3-1	TIME-STRATIGRAPHIC CORRELATION OF THE GLACIAL HISTORY OF THE STUDY AREA	87
4-1	F-TEST RESULTS	172
4-2	T-TEST RESULTS	173

LIST OF FIGURES

FIGURES Page

1-1 RELATIONSHIP BETWEEN MEAN ANNUAL RUNOFF AND PRECIPITATION 6

1-2 RELATIONSHIP BETWEEN MEAN ANNUAL SEDIMENT YIELD AND PRECIPITATION 7

1-3 RELATIONSHIP BETWEEN MEAN ANNUAL SEDIMENT CONCENTRATION AND PRECIPITATION 8

1-4 CHANGES IN STREAM EQUILIBIRUM GRADE 11

1-5 DRAINAGE BASIN RELIEF AND RATES OF UPLIFT AND DENUDATION .15

1-6 GLACIAL FOREBULGE 16

1-7 CHANGES IN STREAM EQUILIBRIUM DUE TO BASE LEVEL CHANGES 18

1-8 CHANGES IN CHANNEL CROSS-SECTION DUE TO BASE LEVEL CHANGES 20

1-9 FINING UPWARD SEQUENCE OF A MEANDERING RIVER. . .24

1-10 MORPHOLOGY AND DEPOSITIONAL FACIES OF A MEANDERING RIVER 25

1-11 TYPES OF BRAIDED RIVER DEPOSITIONAL PROFILES. . .30

1-12 COMPARISON BETWEEN BRAIDED AND MEANDERING VERTICAL PROFILES 33

1-13 TWO SEQUENCES OF EVENTS LEADING TO TERRACE DEVELOPMENT .34

FIGURE Page

1-14 VALLEY CROSS-SECTIONS SHOWING POSSIBLE STRATIGRAPHIC RELATIONS IN VALLEY ALLUVIUM. . . .36

1-15 ALLUVIAL TERRACE STUDIES IN ALBERTA: SITE LOCATIONS .38

1-16 VALLEY TRAIN TERRACE GRADIENTS ALONG THE OLDMAN RIVER, ALBERTA 40

1-17 TERRACES NORTH OF THE BOW RIVER, COCHRANE, ALBERTA .46

1-18 TERRACE DEVELOPMENT OF THE RED DEER RIVER, ALBERTA .51

1-19 LOCATION OF WHITEMUD AND WEED CREEKS, ALBERTA . .56

2-1 ATHABASCA RIVER BASIN 62

2-2 ATHABASCA RIVER STUDY AREA 63

2-3 ATHABASCA RIVER LONG VALLEY PROFILE 65

2-4 GENERAL CROSS-SECTIONS OF THE ATHABASCA RIVER . .66

2-5 HYPOTHETICAL MODEL OF TERRACE GRADIENTS IN RESPONSE TO EPISODIC BASE LEVEL LOWERING 70

2-6 POSSIBLE TERRACE RELATIONSHIPS ALONG THE ATHABASCA RIVER 71

2-7 STREAM FLOW DATA FOR THE ATHABASCA RIVER (1974) .77

2-8 EFFECT OF SUMMER PRECIPITATION ON RIVER DISCHARGE (1969) 78

2-9 DIFFERENCES IN DISCHARGE OF THE ATHABASCA RIVER AT HINTON AND WINDFALL (1977) 80

FIGURE Page

2-10 DIFFERENCES IN DISCHARGE OF THE ATHABASCA RIVER AT HINTON AND WINDFALL (1961-1974) 81

2-11 GEOLOGY OF THE STUDY AREA 82

2-12 GEOLOGICAL CROSS-SECTION AA'. 85

2-13 GEOLOGICAL CROSS-SECTION BB'. 85

3-1 LOCATION OF SURFICIAL GEOLOGY MAPS: FIGURES 3-2 to 3-7. 88

3-8 CORDILLERAN/LAURENTIDE ICE CONTACT LINE 102

3-9 GLACIAL LAKE WINDFALL 106

3-10 GLACIAL LAKE WILDWOOD 107

3-11 GLACIAL LAKE LEDUC. 108

3-12 C^{14} DATES IN THE PROXIMITY OF THE ATHABASCA RIVER 110

4-1 ELLIPTICAL TRACE. 134

4-2 SAMPLE MEAN OF APPARENT AXES. 136

4-3 CORRESPONDING TRUE AXES 136

4-4 LOCATION OF TERRACE MAPS: FIGURES 4-5 to 4-10 . 139

4-11 VALLEY LONG PROFILE SHOWING TERRACE REMNANTS. . 152

4-12 VALLEY CROSS-SECTIONS: LINE 1 to LINE 8 154

4-13 VALLEY CROSS-SECTIONS: LINE 9 to LINE 14. . . . 156

4-14 COMPOSITE DIAGRAM OF CROSS-VALLEY PROFILES. . . 157

4-15 FREQUENCY DISTRIBUTION OF GRAIN SIZE CALCULATIONS. 169

5-1 IDEALIZED VALLEY PROFILE. 180

LIST OF PLATES

PLATE		Page
2-1	QUASI-BRAIDED CHANNEL WITH SPOOL BAR AND POINT BAR COMPLEX.	64
2-2	IDENTIFICATION MARKER OF POTENTIAL HAZARD ENCOUNTERED IN THE STUDY AREA.	73
3-1	ILLUSTRATED MAYBERNE TILL SECTION 6-5-2.	112
3-2	ILLUSTRATED MAYBERNE TILL SECTION 7-31	112
3-3	MAYBERNE TILL OVERLYING BURIED VALLEY GRAVELS, SECTION 6-12	113
3-4	MARLBORO TILL OVERLYING BEDROCK, SECTION 3-3 . .	115
3-5	ILLUSTRATED MARLBORO TILL SECTION 3-4.	115
3-6	ILLUSTRATED OBED TILL SECTION 1A-5	117
3-7	ILLUSTRATED OBED TILL SECTION 1A-1	117
3-8	PEDLEY SEDIMENTS OVERLAIN BY OBED TILL	119
3-9	ILLUSTRATED PEDLEY SEDIMENTS SECTION 1A-4. . . .	120
3-10	ILLUSTRATED SECTION 1-1 OF T_{2H} ALLUVIUM.	122
3-11	ILLUSTRATED SECTION 1-1 LOWER ALLUVIAL UNIT. . .	122
3-12	ILLUSTRATED SECTION 1-1 UPPER ALLUVIAL UNIT. . .	123
3-13	ILLUSTRATED BERLAND OUTWASH SECTION, 6-33. . . .	123
3-14	ILLUSTRATED DELTAIC SAND SECTION 7-8	125
3-15	RELICT DUNE FIELD.	125
3-16	LOW LYING MARSH AREA RESPONSIBLE FOR THE DEPOSITION OF ORGANIC DEPOSITS	127

PLATE Page

4-1 EXAMPLE OF GRID PLACEMENT FOR GRID-BY-NUMBER SAMPLING TECHNIQUE. 132

4-2 DISTINCT CONTACT BETWEEN TERRACE ALLUVIUM AND UNDERLYING BEDROCK. 161

4-3 FINING-UPWARD OF ALLUVIAL GRAVELS, T_2 161

4-4 ILLUSTRATED TERRACE TWO SECTION 6-7 162

4-5 CONTACT BETWEEN T_3 ALLUVIUM AND UNDERLYING BEDROCK 164

4-6 ILLUSTRATED TERRACE THREE SECTION 7-30. 164

4-7 AERIAL PHOTOGRAPH OF REMNANT DUNE FIELD 177

INTRODUCTION

Much of the information on the Quaternary history of west-central Alberta has focused on the delineation of glacial deposits (Collins and Swan, 1955; Henderson, 1959; Roed, 1968, 1975), the location of past ice frontal positions, and the sequence of proglacial laké deposits and outlet spillways (Stalker, 1960; Taylor, 1960; St-Onge, 1972). More limited work on the evolution of river valleys, as indicated by alluvial stratigraphies and terrace remnants, has been conducted to assist interpretations of the Quaternary history of the region. The effects of the Cordilleran and Laurentide ice masses have been interpreted largely on the basis of stratigraphic evidence, particularly that related to the characteristics of exposed tills.

For two main reasons the Athabasca River valley is of key importance in the interpretation of the Quaternary history of west-central Alberta. First, the Athabasca River provides an excellent area in which to link together earlier studies by Roed (1968, 1975), and Stene (1966) in the west, with that of St-Onge (1972), to the east. Second, the variation of alluvial stratigraphies and terrace units in the river basin may be correlated with reasonably well-known glacial events of Laurentide and Cordilleran origins.

The four main objectives of the study were as follows:

1. to determine the distributions and relationships of the present floodplain, terrace remnants and valley-top margins,

2. to log accessible sedimentary exposures, determine the characteristics of terrace alluvium, and to distinguish between Cordilleran and Laurentide stratigraphic units,

3. to relate the distribution of terrace remnants to known levels of proglacial lakes and spillways (described by St-Onge, 1972) and known glacial events in the upper Athabasca River valley (Roed, 1968, 1975; Luckman, 1977; Luckman and Osborn, 1978).

4. to present an integrated interpretation of Late Quaternary glaciofluvial and fluvial activities along the linking part of the Athabasca River valley.

CHAPTER I

ALLUVIAL RIVER TERRACES

1.1 Causative Factors of Alluvial Terrace Development

1.1.1 Introduction

An extensive terminology of fluvial terrace morphologies is now well established in the literature (see for example, Cotton, 1940; Culling, 1957; Quinn, 1957; Leopold, et. al., 1964; Howard, et. al., 1968; Dury, 1970; Schumm, 1977). Yet, problems remain in relating terrace sequences and alluvial deposits to changing Quaternary environments. The adjustments of hydrologic regimes to Quaternary changes of external variables produced more complex responses than are usually recognized. This complexity would have been compounded for very large drainage basins spanning a wide variety of local environments.

Three external variables are primarily related to terrace development. River terraces, and the episodic patterns of sedimentation revealed by alluvial deposits, are indicative of the long-term adjustments of a basin's hydrologic regime to climatic, tectonic/isostatic, and/or base level fluctuations. The general effects of these controls on hydrologic and hydraulic regimes were usefully demonstrated by Lane (1955). He adopted a qualitative relationship for stream equilibrium

based on the quantity of sediment (Q_s), the particle size or diameter of the sediment (d), water discharge (Q_W), and slope (S), of a stream:

$$Q_s d \simeq Q_w S \qquad [1]$$

A change in one of the hydraulic variables, in response to external controls, resulted in the alteration of drainage basin regimes. Such hydraulic responses, in turn, stimulated episodes of valley filling and cutting, with transitional phases of river equilibrium.

1.1.2 Climatic Controls

Climatic control of the hydrologic regime is indirect. The causes of river terraces whose origins can be traced to a change in climate, rather than diastrophism, are highly complex (Schumm, 1965). Flint (1957, p. 218) explained that;

> Climatic inferences from alluvial stratigraphy are difficult because of the number of variable factors involved. Any climate consists of a group of variables such as the amount of precipitation, distribution of precipitation throughout the year, mean and seasonal temperatures, and the like. The response of a stream, in terms of discharge and load, to a change in one or more of the climatic variables will be affected by local topographic texture, steepness of slopes, character of vegetation cover, and other circumstances. Hence a change in only one climatic factor might lead to very different responses in two different streams, and even in two different segments of a single long stream.

To illustrate the effects of climatic change on sediment yields Langbein's et. al., (1949) rainfall/runoff curves for

temperatures of 4.5°C (40°F), 10°C (50°F), 15.5°C (60°F) and 21°C (70°F), (Figure 1-1), were adopted by Schumm (1965) to derive the sediment-yield curves shown in Figure 1-2. While precipitation is probably the dominant climatic factor influencing sediment yields the effects of temperature must also be considered. More precipitation is required to produce a given amount of runoff in a warm climate than in a cold one. The curves show that, within limits, as annual temperatures increase the peak sediment yield occurs at higher amounts of annual precipitation; that is, as temperatures increase, rates of evaporation and transpiration also increase, thereby greater amounts of precipitation are required to maintain the peak rate of sediment yield (Schumm, 1977).

Sediment concentrations are also important. Figure 1-3 shows Schumm's (1965) suggested relationship between sediment concentrations and precipitation at different temperatures. These curves may be used to estimate approximate changes in sediment yields as temperatures, precipitation or both, changed in the past. Only the directions and approximate magnitudes of such changes are meaningful, however, as variations in local geology and geomorphology would also have affected sediment yields (Schumm, 1965).

Schumm (1965, p. 788) cautioned that these curves are applicable only to nonglaciated and/or pluvial regions. In referring to glacial areas he found that;

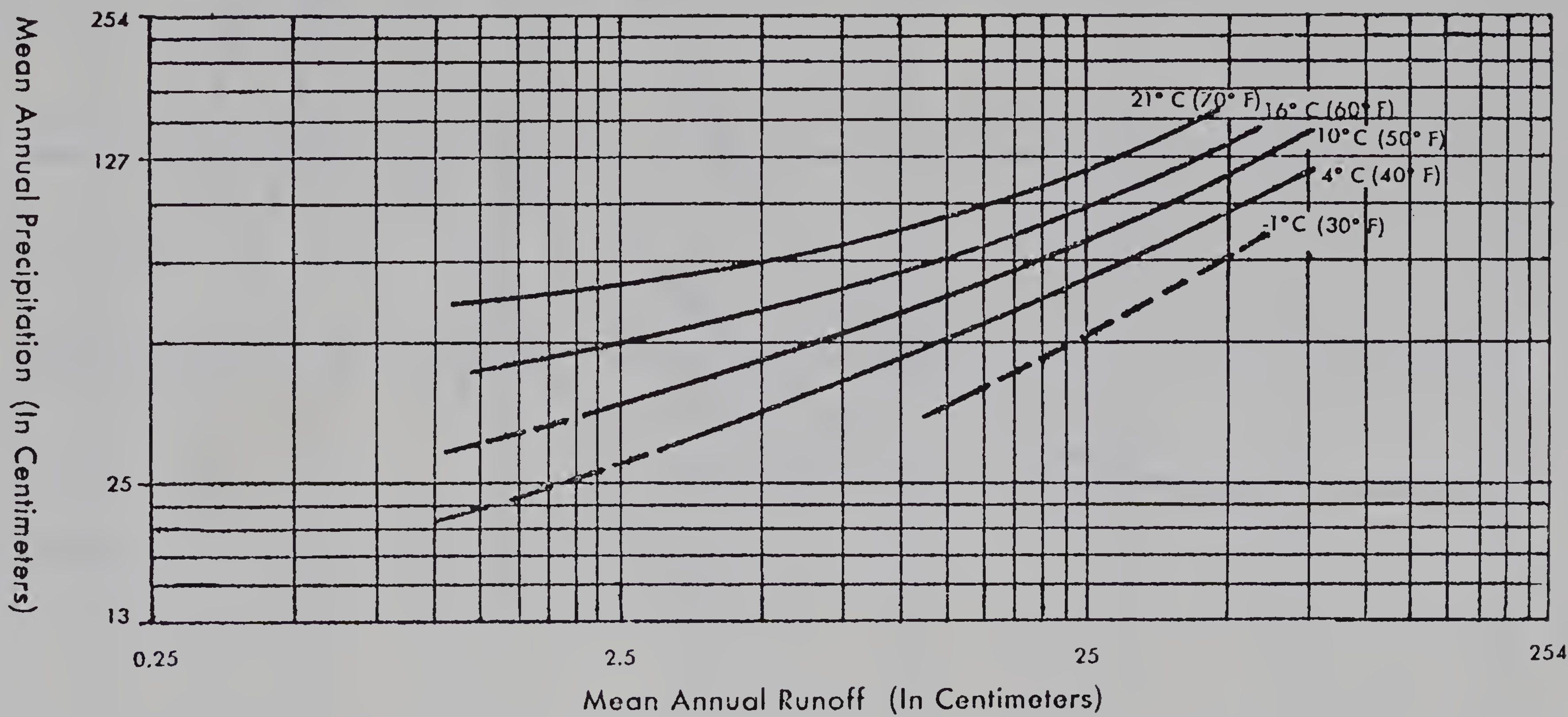


Figure 1-1. Curves illustrating the effect of temperature on the relation between mean annual runoff and mean annual precipitation, (after Langbein et. al., 1949).

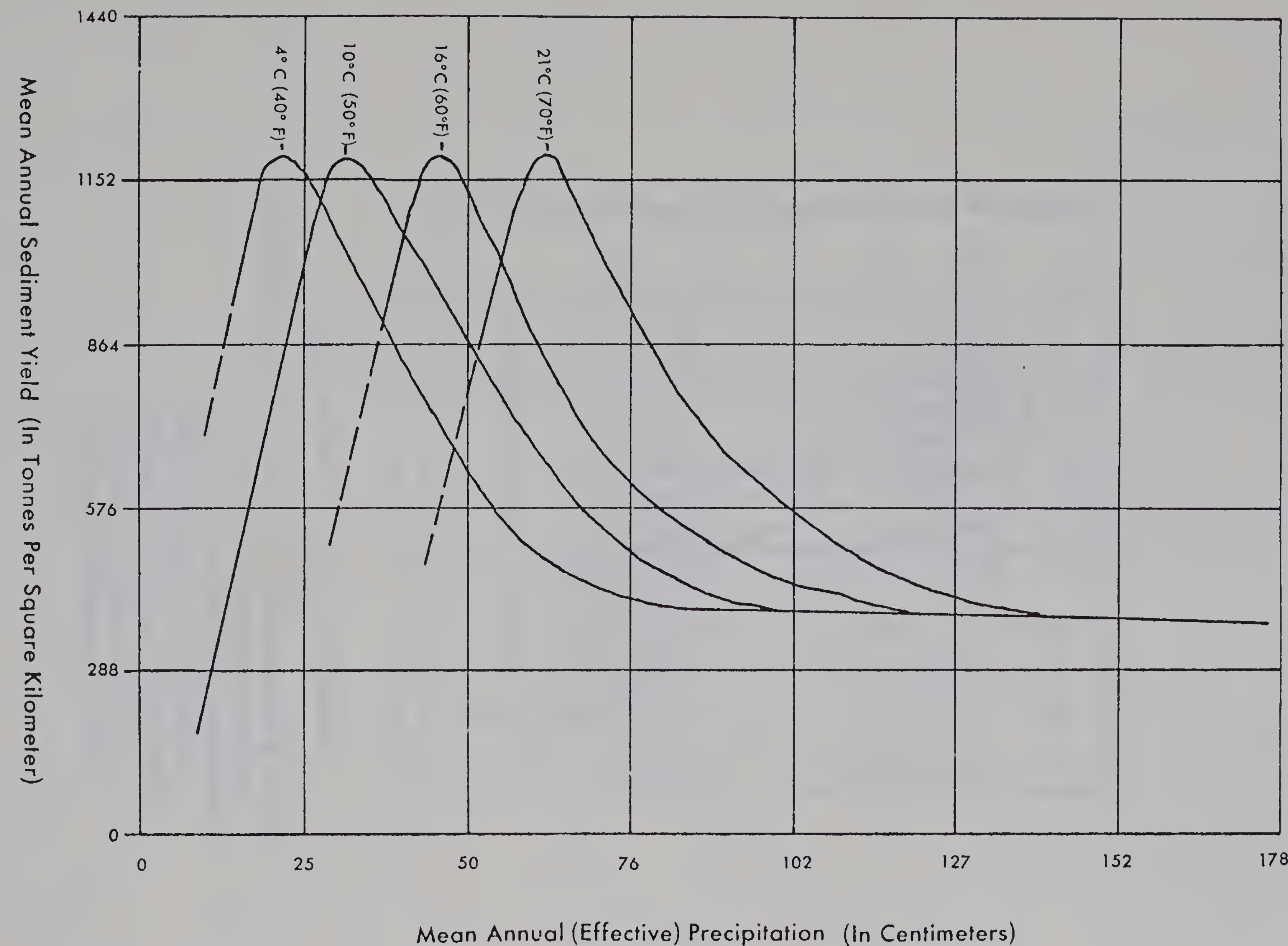


Figure 1-2. Curves illustrating the effect of temperature on the relation between mean annual sediment yield and mean annual precipitation, (after Schumm, 1965).

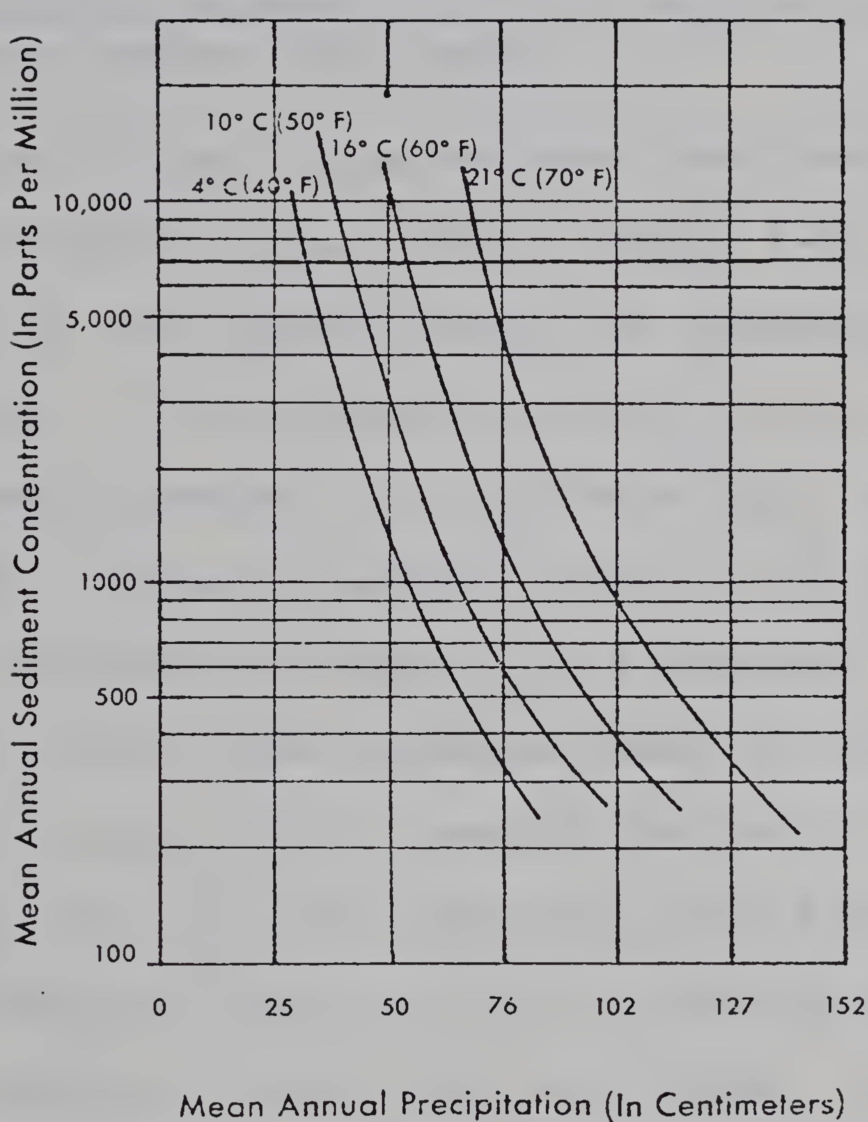


Figure 1-3. Curves illustrating the effect of temperature on the relation between mean annual sediment concentration and mean annual precipitation, (after Schumm, 1965).

For these regions a decrease in temperature and an increase in precipitation will cause an increase in runoff and a decrease in sediment concentration (Table 1A). Sediment yield, however, can either increase or decrease depending on the temperature and precipitation before the change. On the other hand, an increase in temperature and a decrease in precipitation will cause a decrease in annual runoff and an increase in the sediment concentration (Table 1B). Sedimentary yield can either increase or decrease depending on the temperature and precipitation before the change.

Changes in the stream sediment load lead to changes in stream equilibria (Lane, 1955). Figure 1-4A illustrates a stream with an equilibrium grade, BA, reaching base level at A. An increase in the sediment load (Q_s) at C, without changing the sediment diameter (d), discharge (Q_w), or base level, would require a change in channel slope (s) to re-establish the channel equilibrium. When Q_s is increased, the stream is incapable of carrying the increased load, and some of it is deposited downstream from C, causing the bed to aggrade to C'. As deposition continues the stream bed should increase to C''. If the new condition continued for an extended period a new equilibrium grade C'''A might be established. Comparable changes would also result if, instead of sediment discharge (Q_s) increasing at C, there was a reduction in the water discharge (Q_w) at C (Lane, 1955).

Changes in sediment and water discharge would also alter the stream grade above C. The stream would be unable to transport all available sediment through this reach of declining average slope, thus aggrading the stream bed upstream of C

TABLE 1-1

ESTIMATED EFFECTS OF CLIMATIC CHANGE ON HYDROLOGIC VARIABLES

Weighted Mean Annual Temperature (degrees F)		Mean Annual Precipitation (in.)		Ratio of Changed to Present Mean Annual Runoff	Ratio of Changed to Present Mean Annual Sediment Yield	Ratio of Changed to Present Mean Annual Sediment Concentration
Present	Changed	Present	Changed			
A. Change to climate during glaciation						
50	40	10	20	>20	0.6	<0.03
50	40	20	30	5	0.5	0.1
50	40	30	40	3	0.8	>0.2
50	40	40	50	2	0.9	-
60	50	10	20	>20	>2.0	-
60	50	20	30	8	0.4	0.05
60	50	30	40	3	0.7	0.2
60	50	40	50	2	0.9	>0.5
B. Change to climate during interglaciation or postglacial Hypsithermal interval						
50	55	10	5	-	-	-
50	55	20	15	0.2	1.2	>6
50	55	30	25	0.5	1.5	3
50	55	40	35	0.7	1.1	2
60	65	10	5	-	-	-
60	65	20	15	<0.1	0.7	-
60	65	30	25	0.3	1.7	7
30	65	40	35	0.5	1.4	3

Source: Schumm (1965)

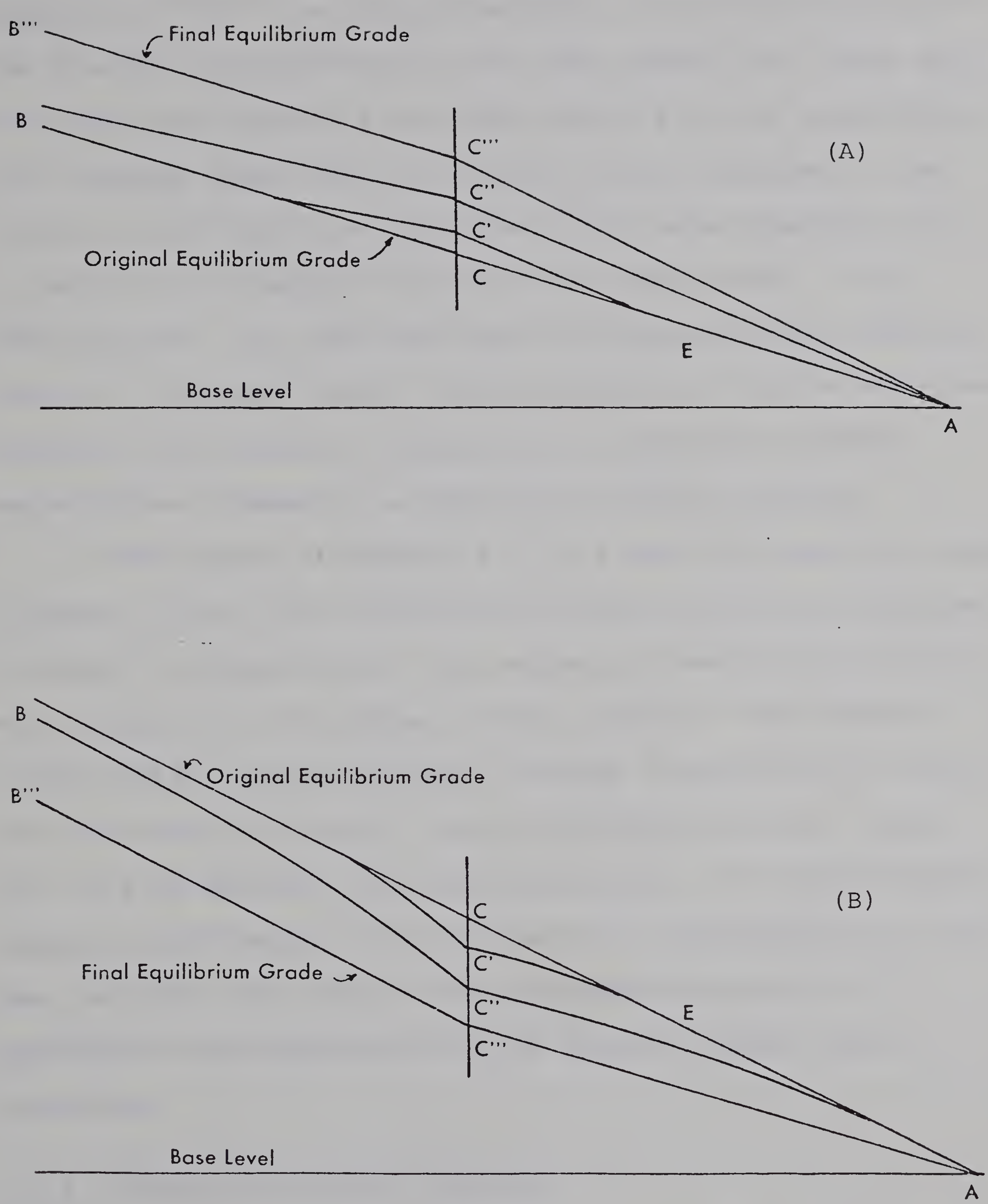


Figure 1-4. (A). Changes in stream equilibrium grade resulting from an increase in sediment load. (B). Changes in stream equilibrium grade resulting from an increase in stream discharge, (after Lane, 1955).

and approaching a long term equilibrium grade, C'''B'''. The height of C'''B''' is largely dependent on the magnitudes of the changes of conditions and the time period over which they influenced the system. A degrading profile of the stream bed, with constant base level, may result from a decrease in the sediment load (Q_s) or an increase in the water discharge (Q_w) at point C, for example (Figure 1-4B) (Lane, 1955). The importance of; (1) lakes and ponds in upstream valley sectors acting as 'sediment sinks' thereby serving to reduce downstream sediment loads; and (2) tributaries in helping to modify master-stream channels, is implicit in these diagrams.

The curves of Figures 1-1, 1-2 and 1-3, which are based on modern runoff and sediment-yield data, suggest the changes in runoff, sediment-yield, and sediment concentrations which may be expected with changes in the climate. The diagrams of the partial stream profiles, Figures 1-4A and 1-4B, suggest the directions of channel slope alteration which may result from varying sediment and water discharges. If these climatic changes significantly alter the sediment-yield/discharge rates over extended time periods the corresponding periods of aggradation and degradation of the channel should lead to terracing.

1.1.3 Tectonic/Isostatic Controls

Tectonic movement of the earht's crust, in the form of either uplift or subsidence, may lead to channel erosion or

deposition respectively, as the gradients of the stream channels in the affected areas are altered by the shifts in the earth's surface. The rate of vertical tectonic movement in some regions of the world can be rapid, occurring at rates capable of influencing a fluvial system over a relatively short period of time. The problems of river stability in these areas are directly related to vertical movements of the crust (Schumm, 1977). It must also be recognized, however, that many areas of the world are undergoing measurable change not only by upwarping or subsidence, but also by lateral displacement. "For example, many small streams are clearly offset along the San Andreas fault of California, where progressive lateral movement on the order of 2.5 cm per year has been measured." (Schumm, 1977, p. 100).

Aside from tectonic movements, alteration of the earth's surfaces may also be attributed to two other processes:

1. isostatic adjustment to denudation, and
2. glacial isostatic adjustment resulting from the removal of temporary loads of glacier ice.

Whether isostatic adjustment to denudation will occur in a given area is dependent on the local strength of the earth's crust such that before isostatic adjustment can occur, this strength must be exceeded by the removal of rock by erosion.

Schumm (1963) illustrated the disparity between rates of uplift and denudation in orogenic areas, suggesting that

isostatic adjustment to denudation is episodic (Figure 1-5). When the initial diastrophism occurred uplift would be relatively rapid. Schumm (1977, p. 60-61) noted that;

> when uplift ceases, denudation proceeds at a relatively slower rate until the strength of the crust is exceeded, when relatively rapid isostatic adjustment should occur again. This relationship is shown diagrammatically in Figure [1-5] where initial uplift is 15000 feet Figure [1-5B]; during and following uplift, denudation rates increase to a maximum, to be followed by a decline as relief is lowered Figure [1-5A]. This sequence of events is interrupted by relatively short periods of isostatic adjustment, during and following which denudation rates again increase to a maximum.

It is generally accepted that large glacial ice sheets constituted a load sufficiently great to cause subsidence of the earth's crust beneath the ice, in a basin-like fashion (Flint, 1977). The stresses caused by the load produced;

1. a depression beneath and extending beyond the actual limits of the ice, and
2. an upward bending of the crust above its equilibrium position to form a forebulge in front of the ice sheet (Figure 1-6).

Because of the stresses produced in the crust by downwarping glacial isostatic rebound is accelerated in the return to equilibrium following deglaciation (Walcott, 1970). Uplift of the area beneath the former ice sheet is relatively rapid, whereas uplift beyond the ice sheet area is slower. Loading by proglacial lakes may also serve to prolong the effects of isostatic depression, by the glacier itself, delaying the earliest phase of crustal recovery.

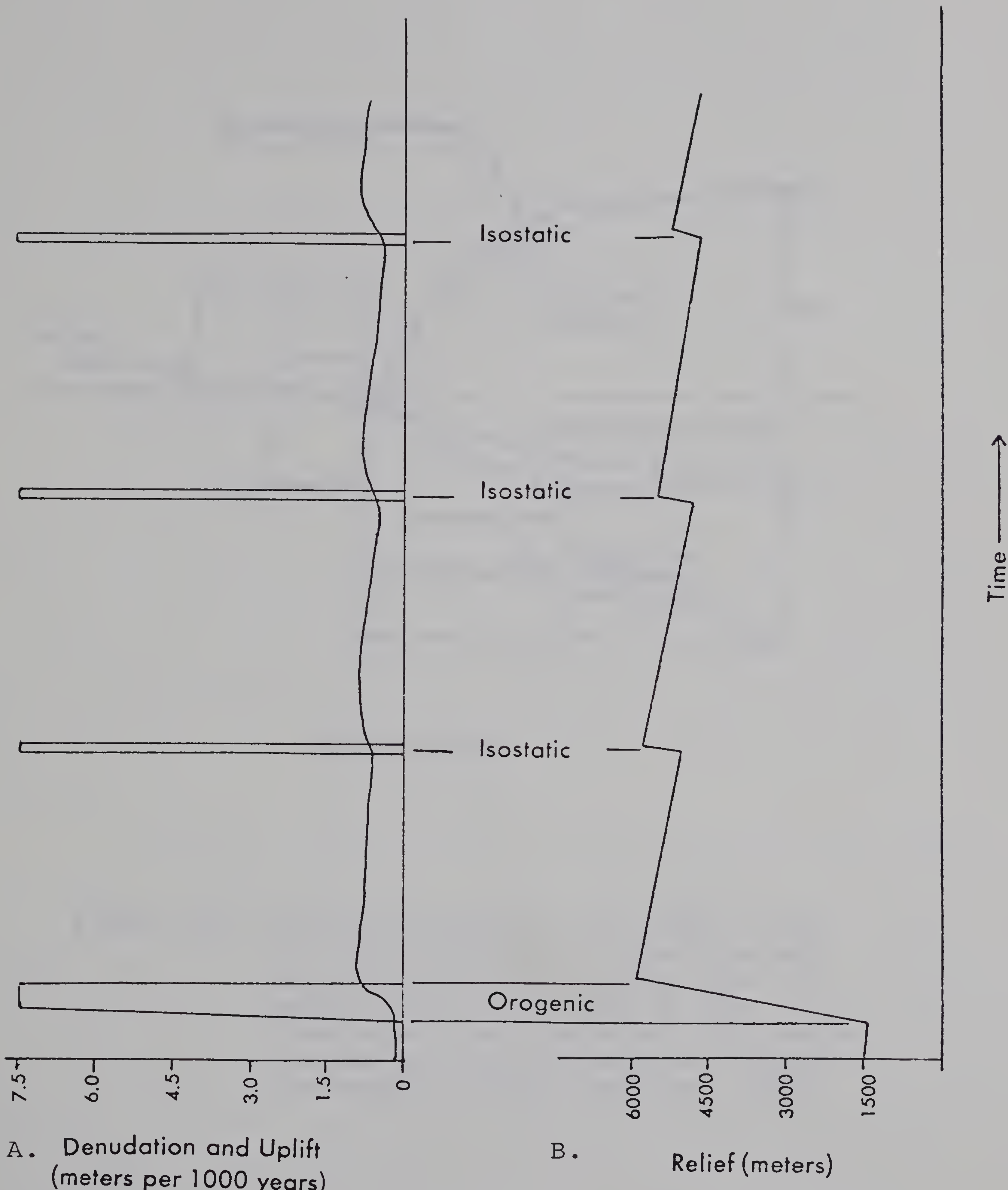


Figure 1-5. A. Hypothetical relation of rates of uplift and denudation (solid line) to time.
B. Hypothetical relation of drainage basin relief to time as a function of uplift and denudation shown in A, (after Schumm, 1963).

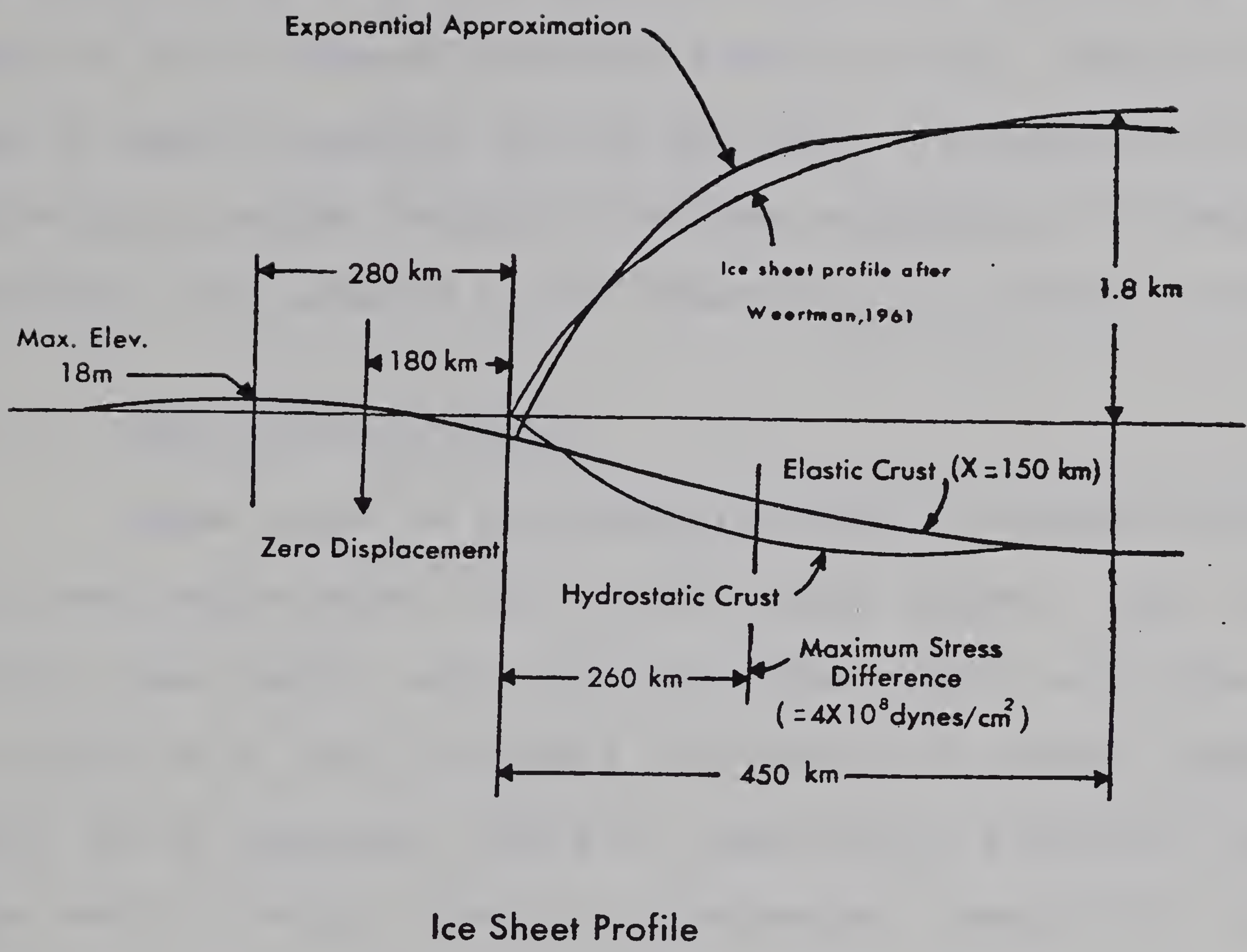


Figure 1-6. Profile across an ice front. The ice sheet profile is approximated by the exponential curve. Note the amplitude and position of the forebulge, the proglacial depression, and the position of maximum stress difference, (after Walcott, 1970).

These two forms of isostatic recovery may be complicated in some areas by tectonic movements. Nevertheless isostatic adjustments operate at greatly differing rates depending on the area in question and the extent of the load exerted on or removed from the earth's crust. This in turn may be used to explain one of the basic alterations in the hydrologic regime leading to stream terracing, as channel gradients are altered by the rebound of the earth's crust.

1.1.4 Base Level Controls

Base level is the downward limit of channel incision, or level below which the channel cannot erode. "The terms 'local base level' and 'temporary base level' are often used to refer to a level to which portions of a channel system flow, or to temporary falls or lakes which similarly may be the level to which rivers or streams may temporarily flow." (Leopold, et. al., 1964, p. 258).

Changes of stream channel profile usually occur when the controlling base level is raised or lowered. The most common cause of a rise in channel base level is dam construction, but this rise may also result from natural causes such as the damming of rivers by landslides, lava flows or glacial advances. In most cases, once drainage through the channel is impeded, a lake then forms upstream of the obstruction, creating a new base level to which the channel will grade (Figure 1-7A). The change in channel gradient which occurs as a result of deltaic deposition in the lake then leads to

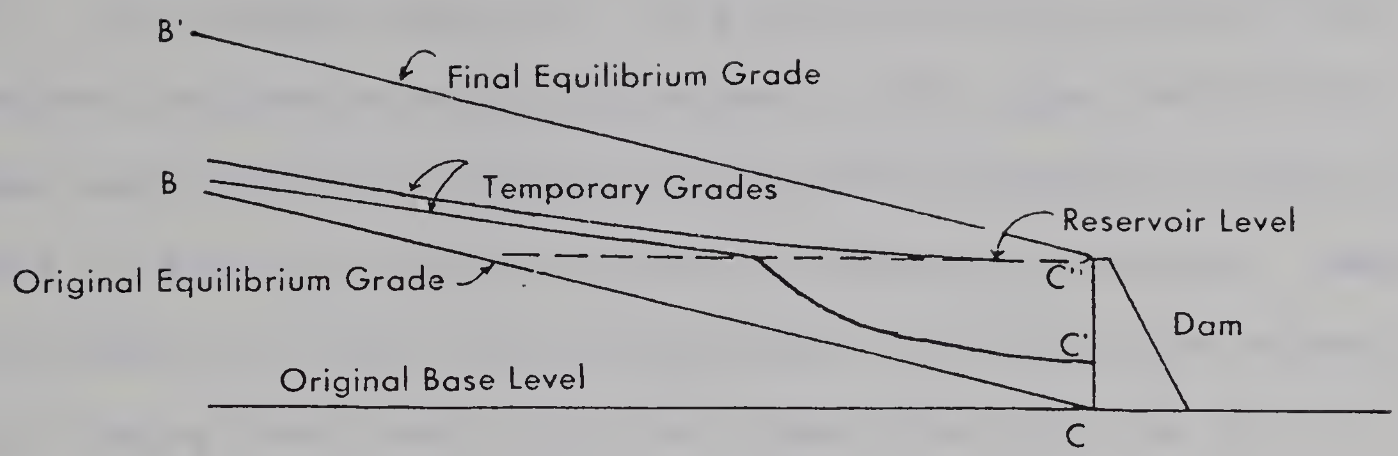


Figure 1-7A. Changes in stream equilibrium grade resulting from a rise in base level. (after Lane, 1955).

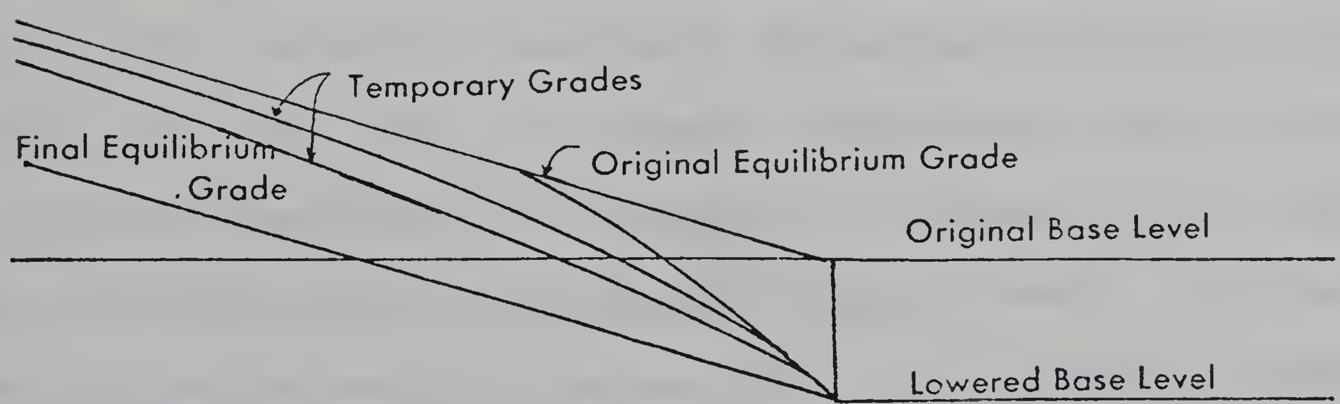
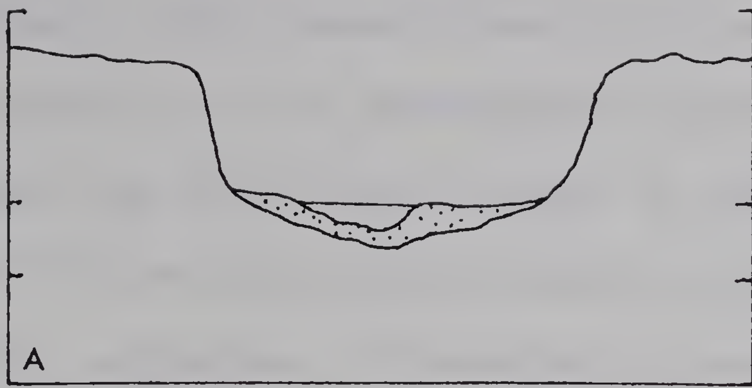


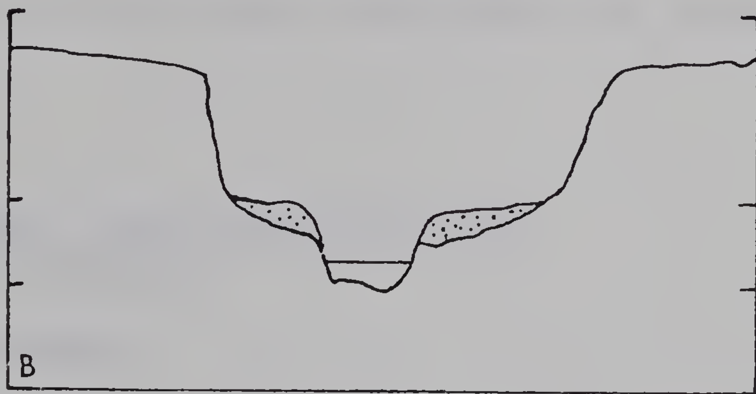
Figure 1-7B. Changes in stream equilibrium grade resulting from a fall in base level. (after Lane, 1955).

progressive upstream deposition in the river, as stream discharge velocities are reduced dramatically as they enter the lake. In Figure 1-7A the grade profile is no longer BC but BC'. The unusual shape of the stream profile results from coarser sediments being deposited at the upper end of the lake and the finer material being carried further into the lake or down to the lake bottom by density currents. Assuming the amount of load to be sufficient to counter the effect of the rising base level, the new base level will be at C''', and the new grade profile will be B'C'''. Aggradation of the original channel helps to re-establish channel equilibrium at the new base level.

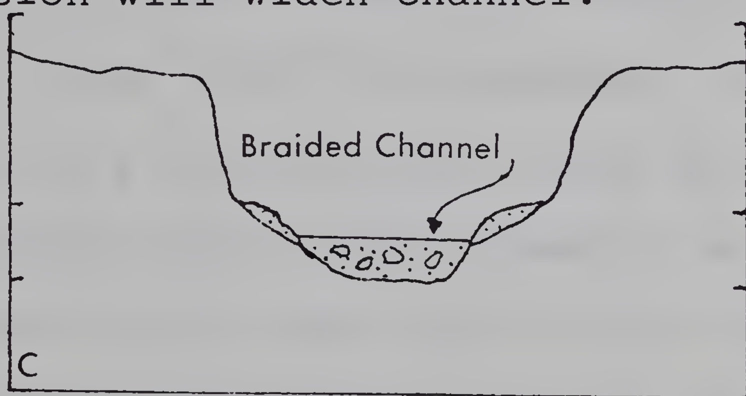
With the lowering of base level the stream trenches previous valley materials, forming one or more terraces. Erosion and channel adjustment to the increased gradient begins near the mouth of the basin and moves progressively upstream (Figure 1-7B), scouring previously deposited valley alluvium (Figures 1-8A and 1-8B). As erosion progresses upstream the main channel experiences an increase in sediment load, leading to the infilling of the new channel (Figure 1-8C). As the stream channel adjusts to the new base level sediment loads decrease and renewed channel incision occurs (Figure 1-8D). Thus initial channel incision, and terrace formation, are followed by deposition of an alluvial fill. As the drainage system again achieves stability renewed incision forms a lower alluvial channel (Schumm and Parker, 1973).



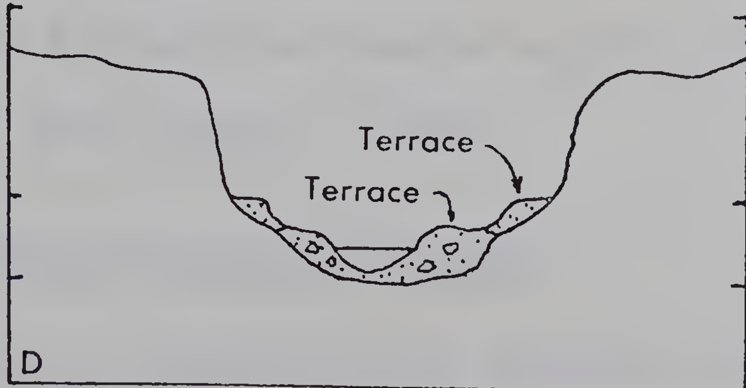
A. Valley before base level lowering.



B. After base level lowering, channel incision forms bedrock terrace. Erosion will widen channel.



C. As sediment yields increase an alluvial fill is deposited and the channel braids.



D. As sediment yields decrease the channel incises into the recently deposited alluvium forming a low alluvial terrace.

Figure 1-8. (after Schumm and Parker, 1973)

In summary, this generalized outline of the causative factors of alluvial terrace development demonstrates that, in natural systems, one event may trigger a complex reaction as the components of the hydrologic regime respond to that event. Alluvial terrace sequences of seemingly identical morphology, may have completely differing origins, as few hydrologic systems reflect the simple influence of a single external variable.

1.2 Sedimentary Characteristics of Alluvial Terraces

1.2.1 Introduction

Terraces are topographical platforms, benches, treads, flats or steps in river valleys, and usually represent former levels of the valley floor or floodplain (Howard, et al., 1968). The continuity of a given surface along the valley and the associated tendency for terrace remnants to occur at a relatively consistent height above the present stream are primary criteria for their correlation (Leopold, et al., 1964). However, for correlations to be adequate and reliable, the stratigraphy of terrace sediments should also be carefully examined (Frye and Leonard, 1954).

1.2.2 Meandering River Stratigraphy

Studies of deposition in channels and on floodplains of meandering rivers show their floodplains to be composed primarily of lateral accretion deposits with an overlay of fine

grained, vertical accretion, flood deposits (Wolman and Leopold, 1957). As a river migrates laterally sediment is deposited below the level of the bankfull stage, on point bars, by lateral accretion. At overflow stages sediments are deposited on the point bars and adjacent floodplain by lateral and vertical accretion. The resulting fining-upward sequences of point bar deposits were described by Allen (1964, 1965) and Visher (1965), among others.

Visher (1965) presented an idealized model of the vertical distribution of sedimentary structures and grain sizes on point bars of sand-bed rivers. The texture decreases upwards in size from poorly sorted, coarse, clastic detritus to very fine sand and silt with some admixture of clay. Sedimentary structures, which reflect an upward decrease in stream energy, are partly related to the changing vertical grain size. From this four fundamental structural units were determined:

1. a basal zone containing poorly sorted coarse material, transported primarily by traction or rolling along the channel bottom.

2. the next zone, characterized by planar cross-bedding, transported by traction and suspension, with bedforms ranging from irregular dunes to rhythmic sand waves,

3. overlying the materials of zone 2 is a zone of horizontally bedded sands, transported by a dense traction carpet moving across a plane bed, exhibiting no ripples or cross-bedding, and

4. the upper unit consists of a symmetrical ripple and/or micro-trough cross-laminated zone, developed from suspension load of the lower flow regime (Visher, 1965).

A fining-upward sequence of point bar deposits, described by Allen (1964), and an interpretation of the morphology and depositional facies of meandering rivers are illustrated in Figures 1-9 and 1-10. The lateral migration of a meander bend erodes the outer bank depositing sediment on the inner banks in a series of ridges (scroll bars) and hollows (swales). The deposit builds laterally in the direction of meander migration, forming a point bar complex. The highest hydraulic energy occurs on the floor of the channel, immediately above the basal erosion surface, where lag concentrations of pebbles and cobbles accumulate. The point bar depositional surface dips gently towards the channel thalweg at angles usually less than 5 degrees. The internal structures typically consist of trough cross-bedding succeeded by ripple marks or climbing-ripples, and cross-lamination of fine-grained sediment, deposited under lower energy conditions. Capping these units are overbank deposits, consisting of silt and clay, with thin sand laminae and carbonate concretions.

1.2.3 Braided River Stratigraphy

Miall (1977) illustrated that various patterns of alluvial deposition may be expressed by braided river deposits depending on, and reflecting variations in, grain size,

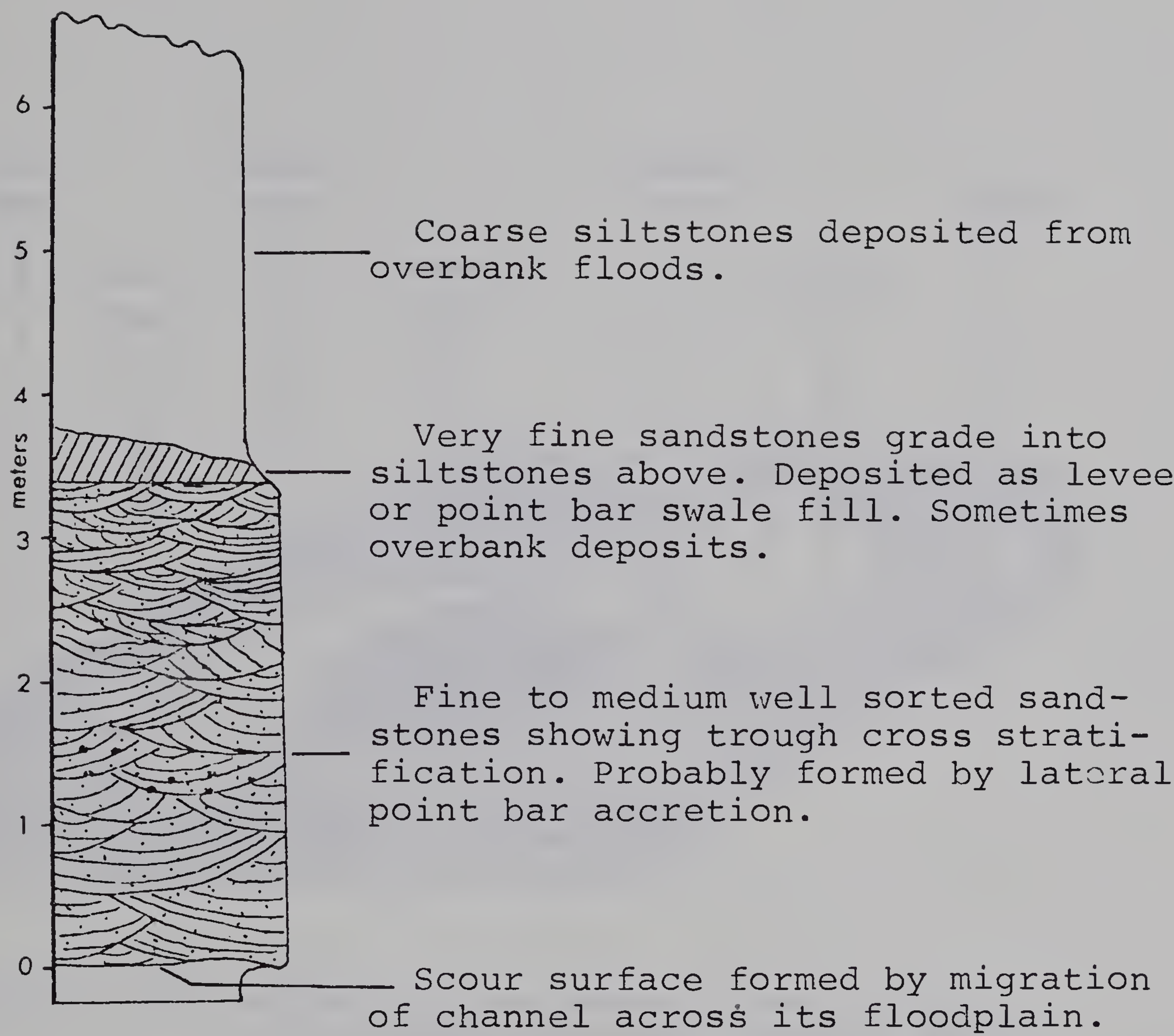


Figure 1-9. Example of a fining upward cycle (generalized), (after Allen, 1964).

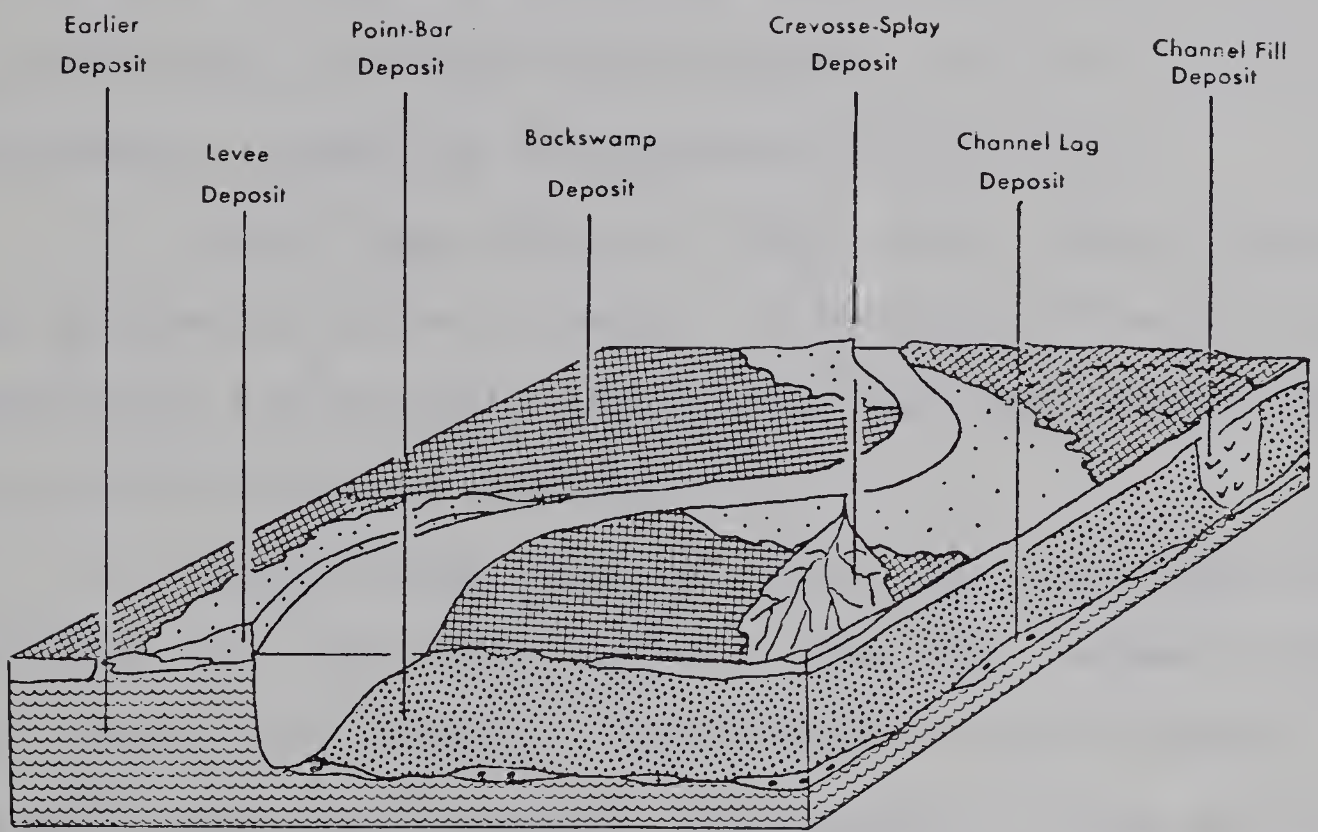


Figure 1-10. Morphology and depositional facies of a meandering river, (Allen, 1964).

discharge variability and rates of aggradation. The principal depositional facies (Table 1-2) of braided streams are interpreted in terms of five main processes (Table 1-3). Each of these processes, except number four, depend on extraordinarily dynamic events, such as extreme peak floods or occurrences of seasonal peak discharges. Miall (1977) derived four idealized sedimentary profile types for braided rivers, any combination of which may be present in given braided stream environments, rendering interpretation difficult:

1. Scott type (Figure 1-11A); this type is characteristic of proximal gravel rivers. It consists of multi-storey, longitudinal bar deposits (Facies Gm) and small scale gravel-sand cycles of waning flood origin.

2. Donjek type (Figure 1-11B); the Scott type grades downstream into this type, which is the most varied of the four. Gravel may compose a major part of the succession, or be absent altogether; several aggradational cycles may be present, of widely varying thickness. All of these cycles originate from an upward decrease in energy levels, as indicated by a decrease in grain size, bed thickness and scale of sedimentary structures in each unit.

3. Platte type (Figure 1-11C); this type represents very shallow, sandy, braided rivers in which gravel is rare or absent. Most of the succession comprises superimposed foreset bar deposits (Facies Sp) with intervals of channel-flow dune deposits (Facies St) and rare overbank deposits (Facies Fl).

TABLE 1-2

Principal Depositional Facies of a Braided Channel

Facies	Description and Origin
Gm: massive or crudely bedded gravels	The facies comprises pebble or cobble gravels in which crude horizontal stratification may or may not be present. Most gravels are clast-supported indicating that the matrix (sand and silt) filtered into the interstices. The facies is common in the braided river environment, in which it occurs as longitudinal bar deposits
Gt: trough cross-bedded gravel	The facies is characterized by broad, shallow channels. The facies is uncommon. It represents scour and fill processes occurring in pebbly, braided rivers.
Gp: planar cross-bedded gravel	The facies results from the migration of linguoid or transverse bars in pebbly braided rivers. Planar cross-bedded sets ranging from 25 cm to 4 m in thickness, and composed of coarse materials, are present. Sand facies include clasts ranging from very fine to very coarse, possibly pebbly, in grain size.
St: trough cross-bedded sand	Solitary or grouped trough cross-bedded sets characterize this facies. Grain size is generally medium to very coarse sand. The cross-beds result from migration of dunes, and are equally common in braided and meandering rivers.
Sp: planar cross-bedded sand	Grain size is medium to very coarse sand. Foresets are avalanche slopes, dipping at 15 to 35 degrees. The facies forms by migration of foreset bars and sand waves, common in sandy braided rivers.

TABLE 1-2 (Continued)

Facies	Description and Origin
Sr: ripple cross-laminated sand	A variety of asymmetrical ripple types, including climbing ripples, characterises this facies. Grain size is very fine to coarse. Occurs in braided or meandering rivers, as low energy bar-top deposits, and as waning flood deposits.
Sh: horizontally bedded sand	Sand may be laminated to massive, fine to very coarse grained. Parting lineations are characteristic. This facies is common as a flood deposit in ephemeral streams, but also occurs as a low energy deposit in other environments.
Ss: scour-fill sand	A rare facies of uncertain origin. Shallow,commonly asymmetric scours occur in sandy braided rivers. The scour fill may contain a variety of internal structures but, unlike the formation of scours in the lee of dunes, the scour and fill events are demonstrably not simultaneous events.
Fl: laminated sand, silt and muck	Sand is rarely coarser than very fine grained. Inter-bedding of sand, silt and muck on a fine scale is common. Very small ripple marks, undulating bioturbation and coal may be abundant. The facies represents river floodplain deposits on the fill of abandoned channels.
Fm: mud or silt drape	Dark coloured, massive or laminated muck or silt occurs as lenticles a few millimeters to a few centimeters in thick-ness. Desiccation cracks are common features. This facies occurs typically as a drape over underlying beds, its lower surface conforming to the shape of any underlying bedform, such as a ripple train. It represents a period of desicca-tion following a drop in water level.

Source: Miall (1977)

TABLE 1-3

The principal depositional facies of braided streams may be interpreted in terms of the following processes.

Processes	Description
(1) Longitudinal bar formation:	Planar or massive bedded bars. They occur mainly in proximal reaches of braided rivers and are the dominant bedform in river gravels. Bars of this type are the principal mode of origin of Facies Gm (Table 2)
(2) Foresetted Bedform Generation:	Facies Gt, St, Sp and Sh form by the generation of dunes, mega -ripples, sand waves, plane beds and the various types of simple foreset bar. Conditions of formation of the various bedforms depend on flow velocity, grain size and flow depth.
(3) Channel scour and fill:	Major and minor channels form by avulsion during high water stage or bar dissection during falling water conditions. Facies Ss form in this way
(4) Low water accretion processes:	A variety of facies, including Sr, Sh, Fl and Fm form during low water stages. These include infills of minor channels and scour hollows, particularly on bar surfaces; ripple and dune accretion on bar surfaces; development of reactivation surfaces; bar-edge sand wedges formed during falling water conditions; drape deposits in pools.
(5) Sedimentation in overbank areas:	Braided rivers are characterized by large areas of floodplain, but many have abandoned areas with sparse thick vegetation cover, which are covered by water at only the highest flood stages. Small scale current structures, bioturbation roots and caliche nodules may be present. Most of these deposits can be assigned Facies Fl.

Source: modified from Miall (1977)

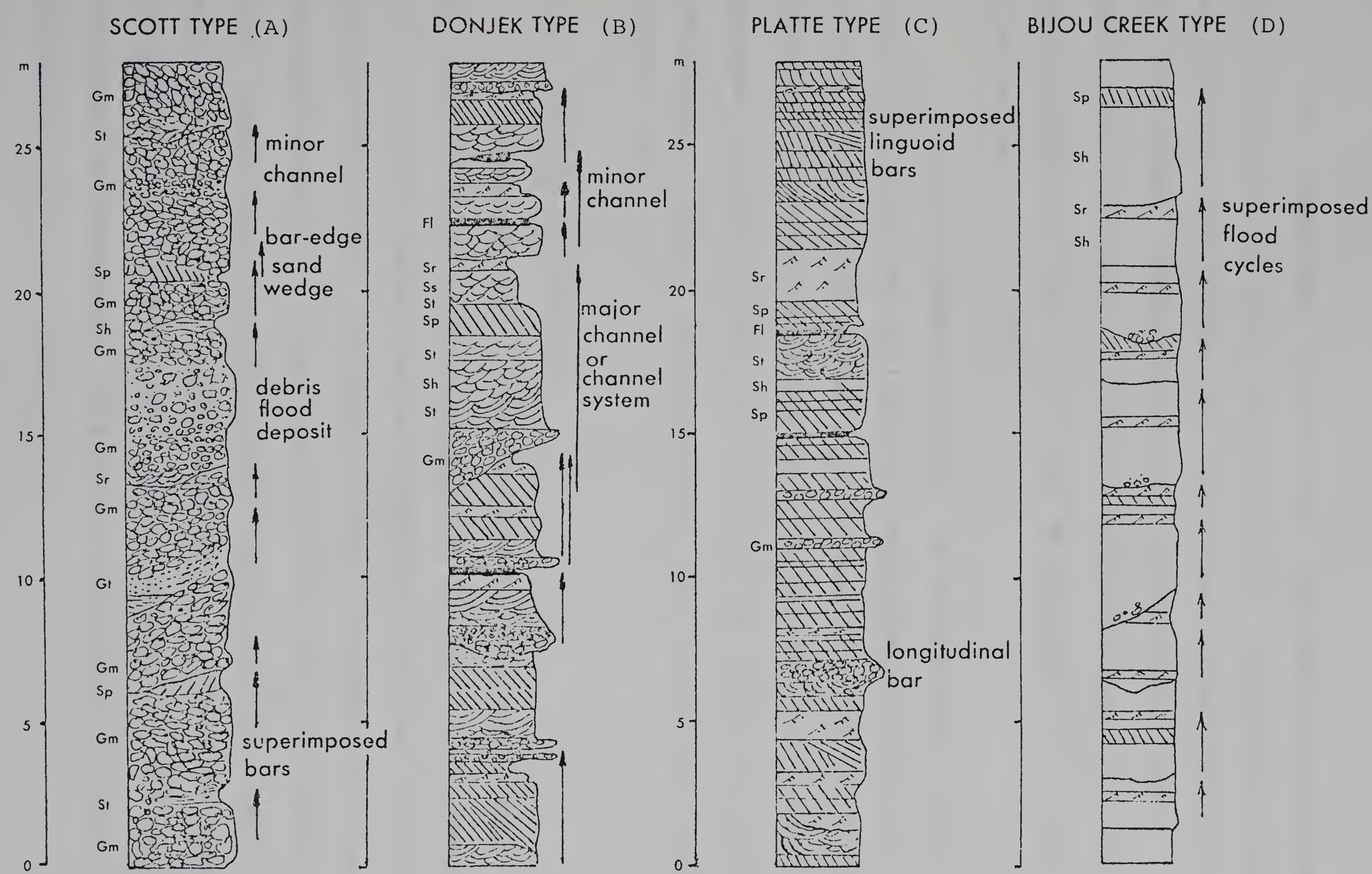


Figure 1-11. Examples of braided river depositional profiles, (after Miall, 1977).

4. Bijou type (Figure 1-11D); Ephemeral sandy streams may be dominated by this type of sediment. Deposition occurs only during flash floods, when upper flow regime conditions may be attained, forming extensive thicknesses of planar bedded sand (Facies St). Facies originating under low flow regime conditions (Sp, St, Sr) may be deposited during waning-flood stages (Miall, 1977).

The within-channel behavior of the principal river types, meandering and braided, are contrasted in Table 1-4. In most instances the sequence of sedimentation is distinctly different. Walker (1976) compared the braided river Donjek sequence, the most easily confused with the meandering fining-upward cycle, with a meandering point bar fining-upward cycle (Figure 1-12).

1.2.4 General Conclusions

Leopold et. al., (1964) defined an alluvial terrace as a former level of the floodplain of a river. The elevation of an alluvial channel changes episodically because of alterations in the channel equilibrium state, triggered by climatic, tectonic/isostatic, or base level fluctuations (see section 1.1). Under such circumstances the floodplain level associated with the previous stream equilibrium is abandoned either by downcutting or aggradation. During downcutting the previous floodplain is dissected, and portions may remain as benches bordering the river (Figure 1-13).

TABLE 1-4

DIFFERENCES BETWEEN THE DEPOSITS OF BRAIDED AND MEANDERING RIVERS

Criteria	Braided Rivers	Meandering Rivers
lateral accretion deposits	point bars, linguoid bars, low water bar accretion	fining-upward point bars (including epsilon crossbeds)
vertical accretion deposits	channel floor bed forms sheet flood deposits bar-top deposits minor overbank deposits	overbank deposits channel lag deposits
type of scour surface	channel erosion	meander widening
channel abandonment behaviour	progressive, as a result of aggradation fill	sudden, as a result of meander neck cut-off
channel abandonment deposit	fining-upward cycle	fine-grained fill
Facies occurrence:		
Gm	common (longitudinal bar deposit)	rare to common (generally as a thin lag deposit)
Gt, Gp	rare to common	absent
St	common	common
Sp	common	generally rare
Sh, Sr	common	common
Ss	rare to common	absent
Fl, Fm	rare to common	common
Channel-fill sequences	rarely > 3m	commonly > 3m

Source: Miall (1977)

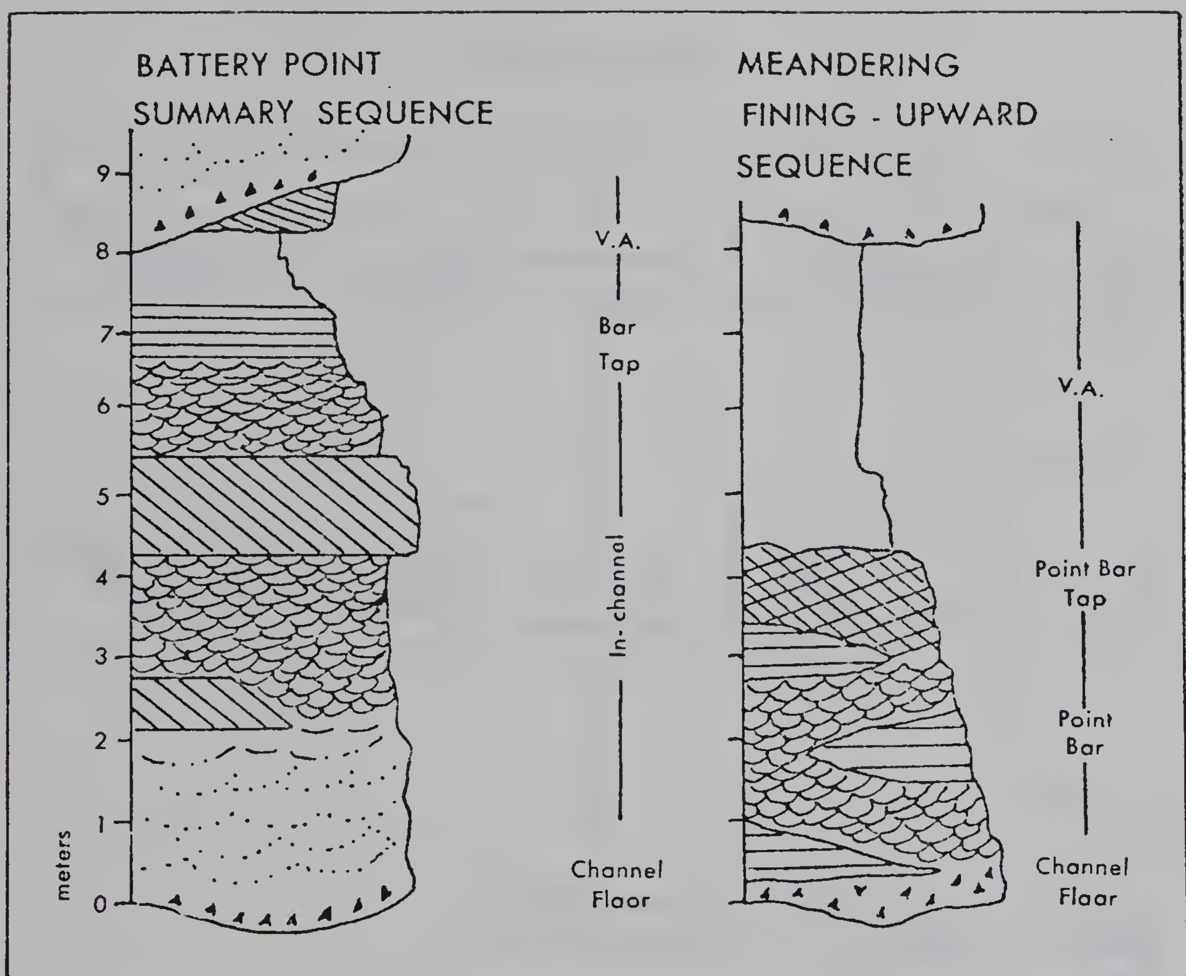


Figure 1-12. Comparison between braided and meandering vertical profiles, (generalized), (after Walker, 1976).

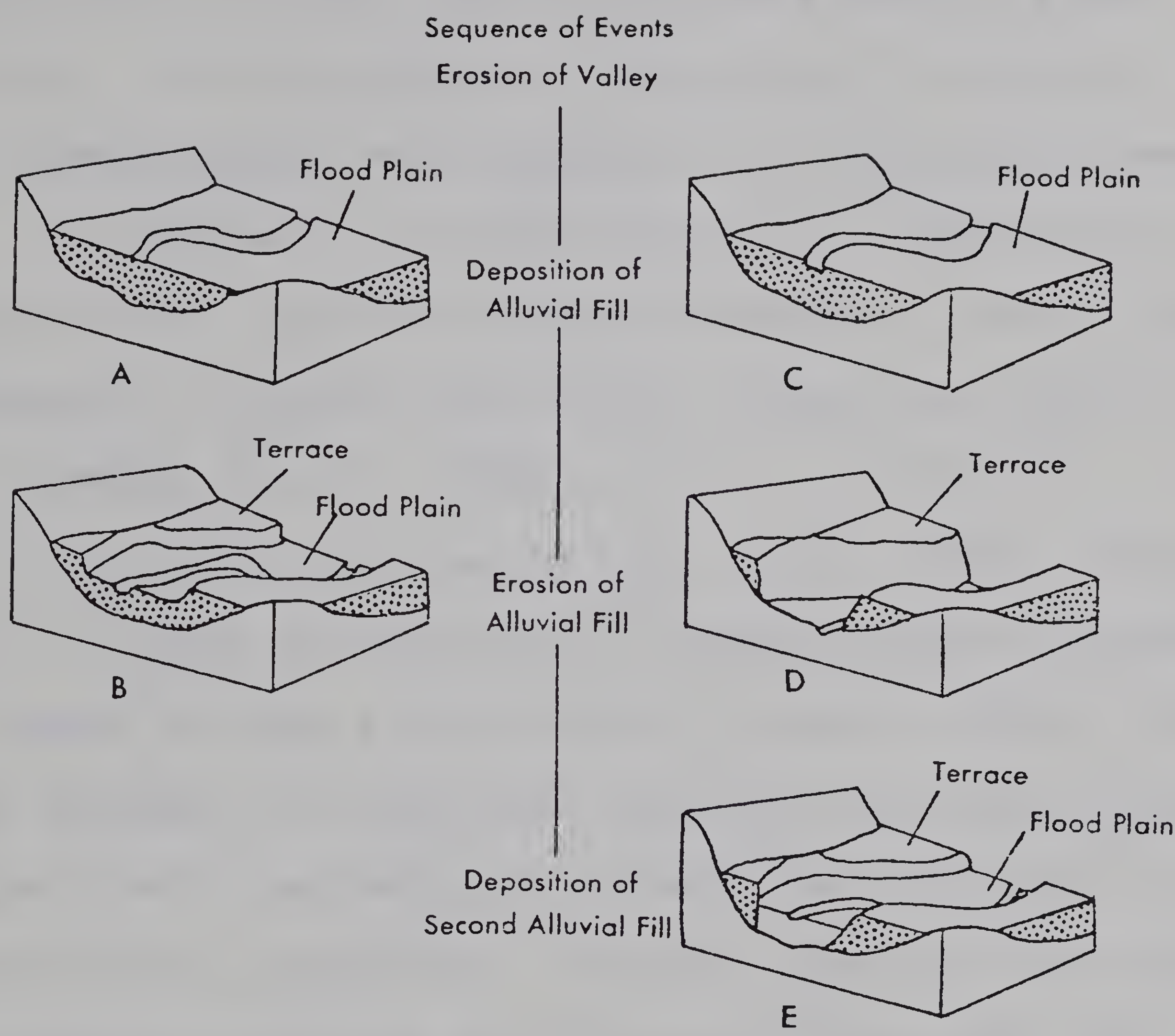


Figure 1-13. Block diagrams illustrating the stages in the development of a terrace. Two sequences of events leading to the same surface geometry are shown in diagrams A,B and C,D,E respectively, (after Leopold et. al., 1964).

When the stream incises below this surface, forming a terrace, downcutting and lateral swinging of the main stream may partly or completely eliminate the original surface. The extent of destruction of the original floodplain surface is variable such that remnant terrace surfaces may be continuous along the valley or occur only as isolated benches along valley walls. Terrace remnants occurring at consistent relative elevations on both sides of the valley are termed 'paired'. In contrast, a stream continually downcutting it's channel may leave a series of isolated terrace treads along the valley sides at differing elevations. These are termed 'unpaired' (Leopold, et. al., 1964).

A distinction is usually made, also, between 'strath terraces', in which the terrace is composed almost entirely of bedrock capped by only a thin veneer of gravels (Moss, 1974), and 'fill terraces' in which the riser and hence the terrace is composed almost entirely of sediment laid down during an earlier period of aggradation. However when alluvium underlies the terrace tread and any portion of the riser, the deposit should probably be refered to as a 'fill' or 'alluvial fill'. In this instance the terms 'strath' or 'fill' terraces becomes confusing and should not be applied (Leopold et al., 1964).

Terraces form an integral part of many valley morphologies and of valley-fill alluvium. Figure 1-14 shows diagrammatically several different stratigraphic and morphological

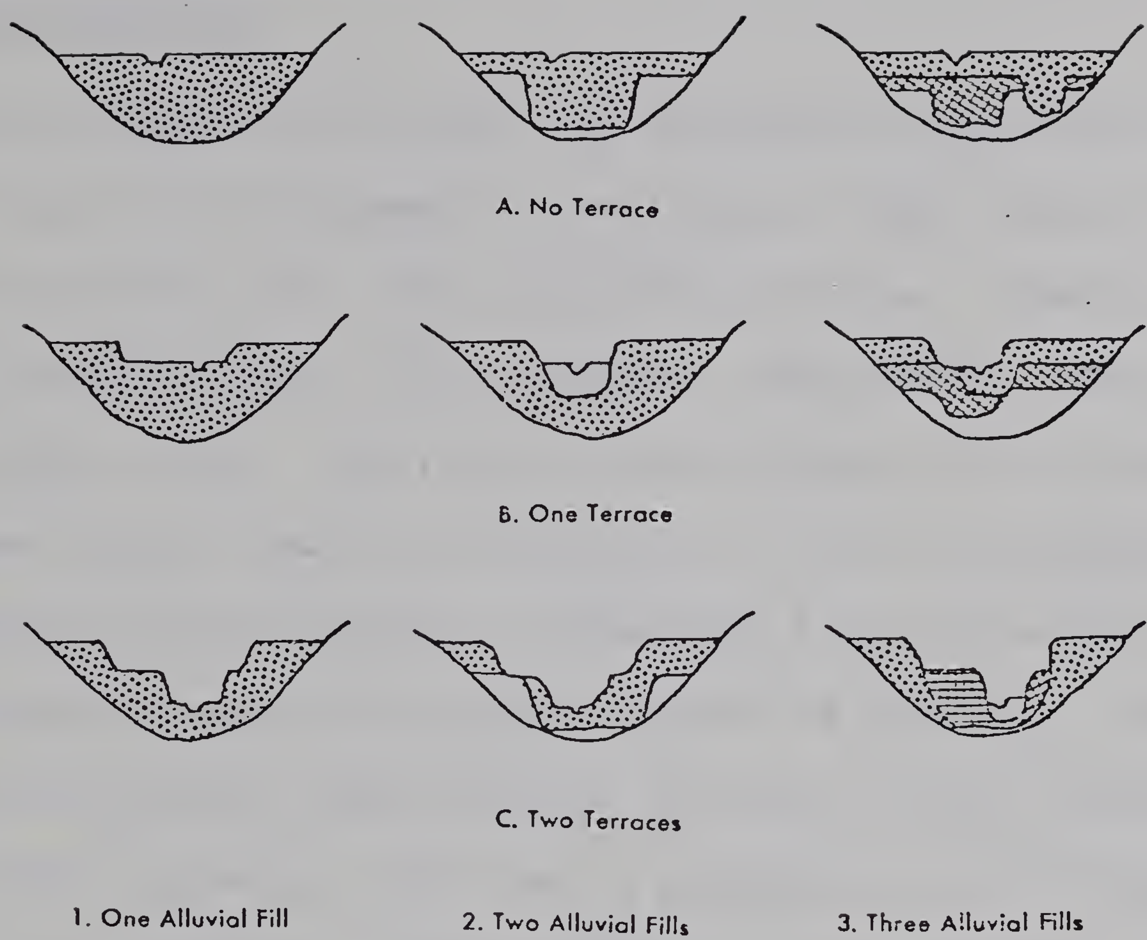


Figure 1-14. Examples of valley cross sections showing some possible stratigraphic relations in valley alluvium, (after Leopold et. al., 1964).

situations which may occur within a valley. The stratigraphic characteristics of the alluvial terrace deposits reflect the modes of alluvial channel deposition prior to the entrenchment of the floodplain to form the terrace. In addition, the morphological characteristics of terraces may demonstrate the complex

responses of the past fluvial regimes to external influences.

1.3 Alluvial Terraces in Alberta: Their Quaternary Significance

1.3.1 Introduction

The causative factors and processes responsible for alluvial terrace development in the major river basins of Alberta are related to late Quaternary events. However, many authors (Stalker, 1963, 1973; Wagner, 1966; Roed, 1968, 1975; Rutter, 1972; Alley, 1973; Harris and Waters, 1977; Boydell, 1978) have relied almost exclusively on the stratigraphic relationships of till units to establish the nature and sequence of Quaternary events for specific parts of Alberta. The tracing of former proglacial lake systems (Stalker, 1960; Taylor, 1960; Rutter, 1972; St-Onge, 1972) and a smattering of C^{14} dates (Rutter, 1972; St-Onge, 1972; Stalker, 1977; Jackson, 1979) have been further sources of material for the interpretation of local Quaternary histories.

Interpretations based on alluvial terraces and alluvial stratigraphies have been afforded lesser prominence in the literature on the Quaternary of Alberta. The following outlines some of the studies which have attempted to correlate alluvial terrace development with major Quaternary events in various regions of Alberta (Figure 1-15). The extent of information and interpretation in these studies is highly variable, many of the studies being much too simplistic in their approach, with the processes responsible for terrace development left

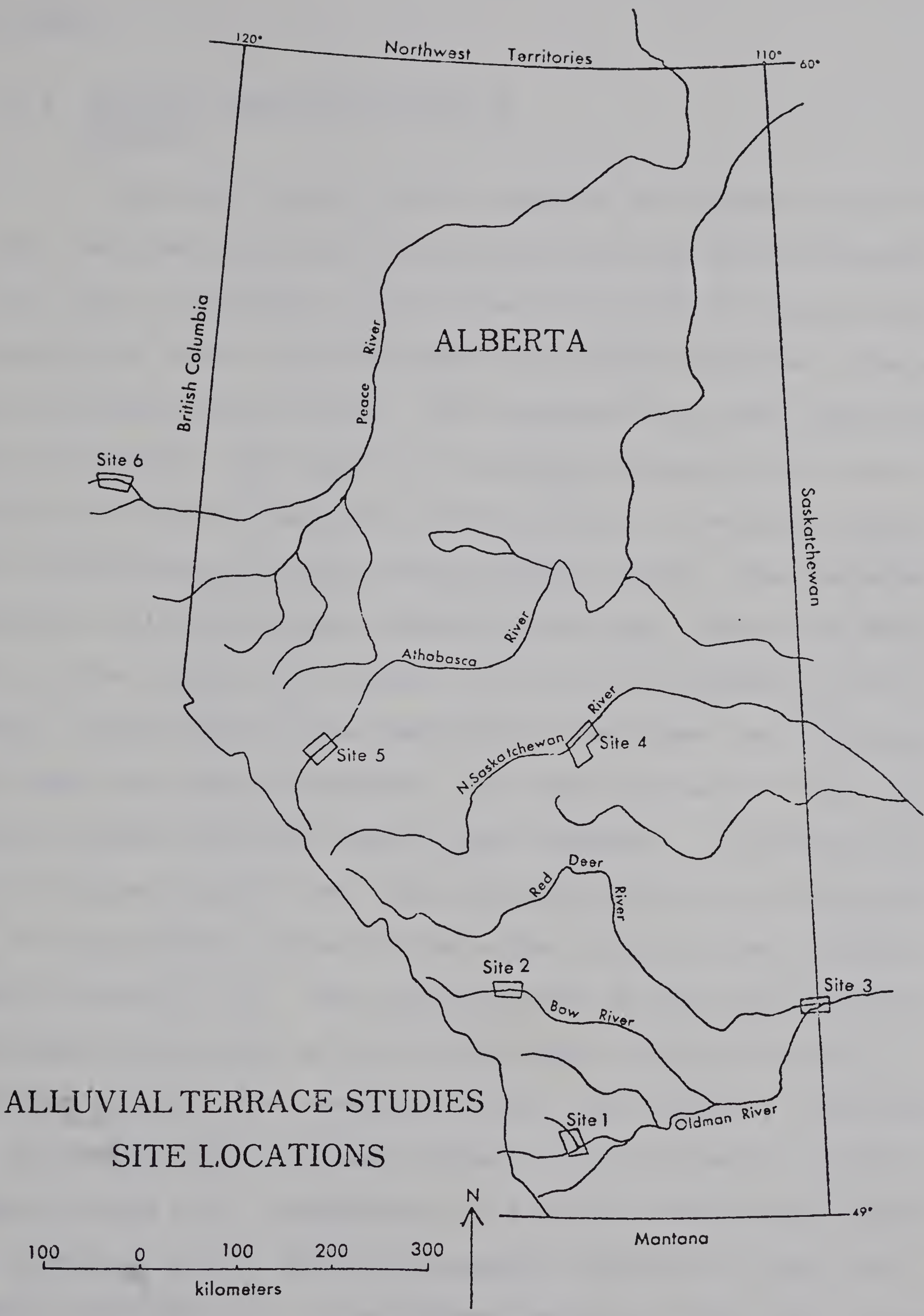

120°
Northwest Territories
110°
60°
ALBERTA
British Columbia
Peace River
Saskatchewan
Site 6
Athobasca River
Site 5
N.Saskatchewan River
Site 4
Red Deer River
Site 2
Bow River
Site 3
Site 1
Oldman River
49°
Montana
ALLUVIAL TERRACE STUDIES
SITE LOCATIONS
100
0
100
200
300
kilometers
N

Figure 1-15

in doubt.

1.3.2 Alluvial Terrace Studies in Alberta

Alley and Harris (1974) examined two flights of valley train terraces along the Oldman and Crowsnest Rivers (Figure 1-15, Site 1), thought to have been deposited during an early recessional stage of Cordilleran ice during the Ernst advance. In the Oldman River valley, ten terrace units were identified by the authors. The upper six terraces terminate upstream in pitted ice-contact deposits, representing the maximum extent of a Cordilleran glacier advance in the region. The proximal terrace surfaces are also marked by kettles, varying in depth from a few meters to 15 meters, and up to 100 meters in diameter. The highest three terraces of the upper set originate at these ice-contact deposits, the lower terraces of the set grade further upvalley beyond these deposits. The gradients of the upper terrace set, when extended downvalley, converge at the approximate level of the higher Glacial Lake Caldwell delta (Figure 1-16). The maximum level of the lake formed a temporary base level to which meltwaters issuing from the glacier flowed. The remaining, lower four terraces, separated in elevation from the upper terrace set by 12 meters in the upper valley area, increasing to 18 meters downstream, grade to a second area of pitted ice-contact deposits in the upper valley (Figure 1-16). The highest terrace of this set originates in these ice-contact deposits, the lower three terraces

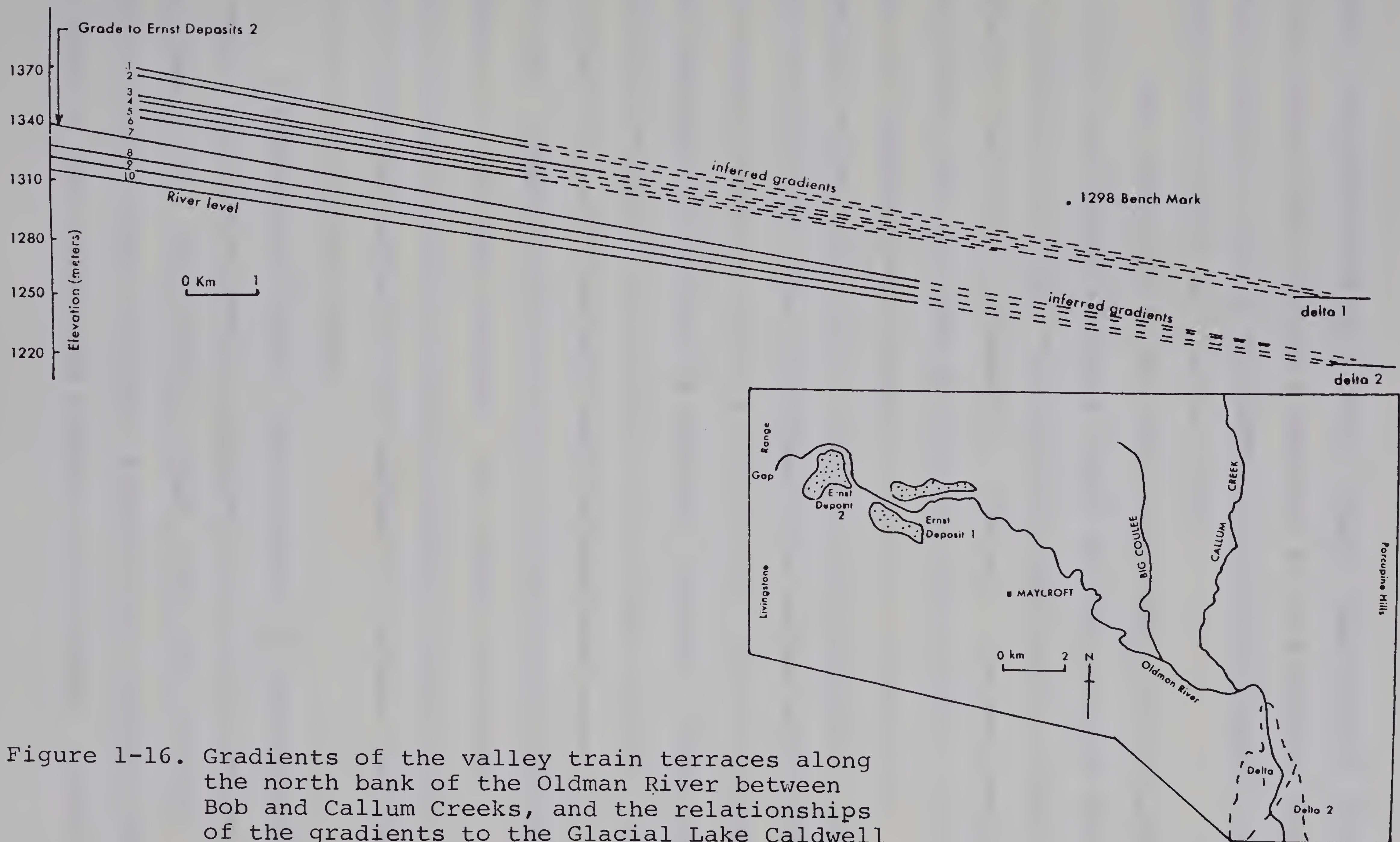


Figure 1-16. Gradients of the valley train terraces along the north bank of the Oldman River between Bob and Callum Creeks, and the relationships of the gradients to the Glacial Lake Caldwell deltas, (after Alley and Harris, 1974).

extend upstream beyond this area. Again, extrapolated gradients of the terrace set converge downstream to an elevation approximately equal to that of the lower Glacial Lake Caldwell delta (Figure 1-16) (Alley and Harris, 1974).

The initiation of the two terrace sets, in areas upstream of pitted ice-contact deposits, would at first suggest that the Cordilleran ice was receding during the period of alluvial deposition and terrace formation. However, a second possible alternative, not discussed by Alley and Harris (1974), is that these are degradational terraces, carved in a previously deposited valley-train fill. Initially, the advance of the Cordilleran glacier down the Oldman River valley would decrease the channel discharge, storing water as ice, but at the same time decrease the distance of sediment transport. Under such conditions accentuated infilling of the valley would be assured. The onset of glacier wastage would release more water into the river, producing downcutting in the previously deposited alluvium, to form the terraces. The pitted ice-contact deposits observed in this area are probably indicative of the transitional period between maximum ice advance and the initiation of glacier retreat.

A lowering of Glacial Lake Caldwell by approximately 30 meters would have greatly increased the channel gradient of the Oldman River. The initial, local, base level lowering would initiate channel incision, erosion being greatest near the mouth of the stream, and progressing upstream. A change in

local base level, with the lowering of Glacial Lake Caldwell, was thus probably responsible for differentiating the two main sets of Oldman River valley terraces (Figure 1-16) (Alley and Harris, 1974).

Two terrace sets in the Crowsnest River valley, further south, illustrate a similar sequence of glacial and proglacial lake influences contributing to terrace development. Alley and Harris (1974), found that:

1. ice-contact deposits marked the maximum extent of the Cordilleran Ernst Advance,

2. downvalley, deltaic deposits may be divided into two units, separated by a difference in elevation of 30 meters,

3. an upper set of six terraces converge downstream to an elevation approximately equal to the upper deltaic unit; the uppermost terrace of the lower terrace unit is coincident with the surface of the lower deltaic unit, and

4. comparable to the Oldman River, the two terrace sets are separated by a sharp break in elevation, illustrating that Glacial Lake Caldwell formed a temporary base level control for the Crowsnest River as well.

In the Bow River valley to the north, at Cochrane, Stalker (1968) identified two terrace sets (Figure 1-15, Site 2). The upper and older terrace set occurs to the north of Cochrane, the lower set to the south of town (Table 1-5). Stalker's (1968) interpretation of the factors responsible for the development of the upper terrace sequence shows some

TABLE 1-5

Terraces north of Bow River at Cochrane, Alberta. Except perhaps for terraces 1 and 2, the higher terraces are older than the lower ones. River low-water level is assumed to be 3670 ft above sea level at Jumpingpound highway crossing.

Reference Number	Approximate height (in ft) above			Composition	Origin	Remarks
	Sea Level	Bow River	Next Terrace			
			Upper Set of Terraces			
1	4000	330	60	largely unknown; medium to coarse gravel (?)	deposit of Bighill Creek into glacial lake	site of Cochrane 'Retreat' may be younger than terrace 2
2	3940	270	30	gravel over bevelled bedrock surface	deposit of Bighill Creek into glacial lake	northern part buried under glacial lake deposits
2a	3910	240	60	thin gravel veneer over bedrock surface	stripping of gravel from terrace 2	subsidiary of terrace 2
3	3850	180	90	medium to coarse gravel	delta built by Bighill Creek into glacial lake	site of High School; flat surface

TABLE 1-5 (Continued)

Reference Number	Approximate height (in ft) above			Composition	Origin	Remarks
	Sea Level	Bow River	Next Terrace			
				Lower Set of Terraces		
4	3760	90	25	coarse gravel	surface of valley fill	site of town of Cochrane
5	3735	65	25	sand to coarse gravel	erosion by Bow River	Clarke Terrace; site of Clarke, Griffin, and other gravel pits
6	3710	40	5	medium to coarse gravel	erosion by Bow River	contains several small gravel pits
7	3705	35	25	medium to coarse gravel	erosion by Bow River	subsidiary to terrace 6
8	3680	10	-	bedrock with veneer of lag gravel	erosion by Bow River	includes modern floodplain and slip-off slopes

Source: Stalker (1968)

similarily to the later interpretations of the Oldman and Crowsnest terraces. The deposits of the upper terrace set at Cochrane (Figure 1-17, numbers 1,2,3) were laid down as deltaic deposits by the Bighill Creek where it flowed into a nearby glacial lake. The creek at that time was much larger than at present, being fed by vast quantities of glacial meltwater. A proglacial lake ponded in front of the Laurentide ice sheet regulated terrace development, forming a temporary base level to which the stream adjusted. Shortly after the formation of the second upper terrace it appears that a Laurentide ice advance raised the local proglacial lake level and thus buried the second terrace beneath 15 meters of lake sediments. These deposits were later stripped from the southern sector of the terrace set but remained on the northern sector. In this instance, the sediments underlying terrace two may in effect pre-date those sediments underlying terrace one. However, incision of the overlying sediments to form the second terrace level would have occurred after the formation of terrace one. Subsequent retreat of the Laurentide ice lowered the proglacial lake level. Bighill Creek re-adjusted to the new, steeper gradient eroding the lake sediments and underlying gravels, until equilibrium was again achieved. Sediments forming terrace three were then deposited. Further recession of the Laurentide ice sheet drained the proglacial

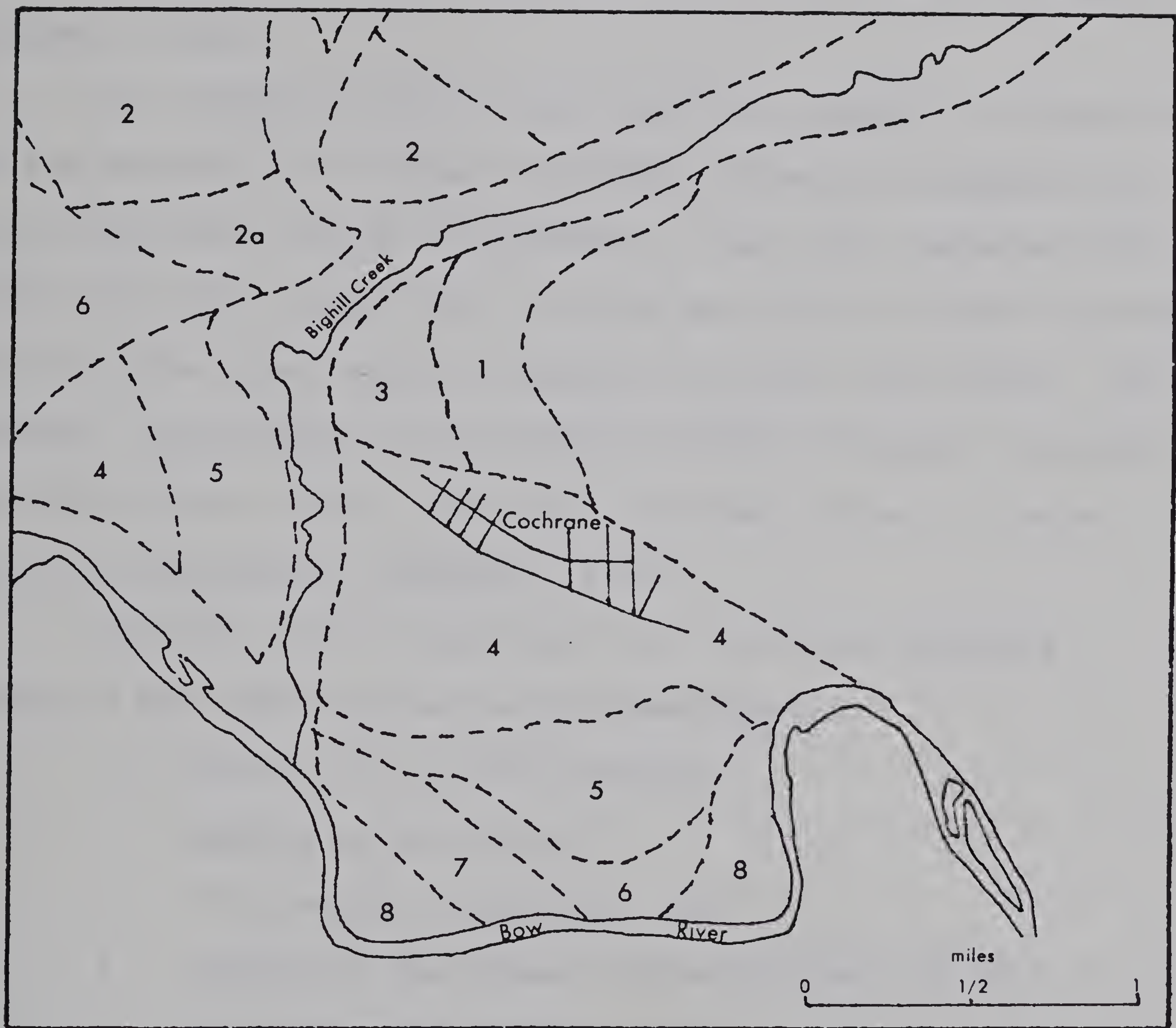


Figure 1-17. Terraces north of the Bow River at Cochrane, Alberta. Terrace numbers and elevations are shown in Table 1-5,(after Stalker, 1968).

lake ending the development of the upper terrace set. The lower terraces (Figure 1-17, numbers 4 to 8), were formed during an extended, postglacial, period of Bow River and Bighill Creek downcutting unrelated to nearby standing water (Stalker, 1968).

The upper terrace of the lower set (number 4) represented the surface of a valley fill which formerly occupied most of the Bow River valley at Cochrane. The other terraces were carved from this valley fill, called the Bighill Creek Formation, after the Bow River began to erode its valley once again. The alluvial stratigraphy of the Bighill Creek Formation suggests that the initial valley fill was laid down during a single period of aggradation (Stalker, 1968).

Stalker (1968) considered the following possible causes of the lower terrace suite development;

1. tilting of the land surface;
2. damming of the river;
3. ice advance or retreat; and
4. changes in the river discharge down the Bow and Kananaskis River valleys.

He immediately eliminated factors one and two. There is no indication of rapid tectonic movement in the area. If postglacial tilting was responsible for deposition it probably resulted from isostatic rebound after the retreat of the glaciers. However, such rebound should have occurred prior

to the formation of the Bighill Creek deposits, for, by the time deposition of the formation is thought to have occurred, the Laurentide ice would have already retreated from most of the prairie region and become much thinner in remaining regions (Stalker, 1968). As well, it is difficult to conceive how rebound could have resulted in 20 meters of alluvial deposition. River damming does not seem possible because the lower terraces continue a considerable distance up and down-valley from Cochrane. In addition, the top of the valley fill (Terrace 4) has a tread gradient similar to the present flood-plain. Local obstruction would have resulted in a more gently sloping tread.

The possible effects of ice advance or retreat on proglacial river behaviour prove to be very complicated in this region. An advance of the Laurentide ice sheet, about the time the Bighill Creek Formation was deposited, would have affected drainage in front of the ice sheet, ponding proglacial lakes and raising the local base level of inflowing rivers. However, at the time of Bighill Creek sediment deposition the Laurentide ice sheet margin may have been 400 kilometers east of Cochrane (Stalker, 1968). If so its effects on river grade at Cochrane would have been negligible. Activities of the closer Cordilleran ice may have been more critical in varying the discharge and sediment load relationships of the Bow River (Figure 1-17, Terrace 4). Stalker (1968) argued that Cordill-eran ice advances would have reduced discharges by storing

water as ice, but would also have decreased the transport distance, enabling more and coarser materials to reach the area. Cordilleran ice retreat would have increased normal stream flows and the supply of materials by releasing water and debris stored in the ice, but would have involved greater distances of sediment transport.

A stable ice-front, upvalley of the Cochrane area, on the other hand, could have had a variety of effects on this area. If the Cordilleran ice stabilized after a retreat period, decreased meltwater flows, but particularly a decrease in sediment discharge, through the valley may have led to river downcutting. However, if the glacier stabilized after a significant advance, as is thought to have happened in the Cochrane area, the primary effect should have been decreased meltwater runoff in the winter, as compared with now, but larger and steadier run-offs during the summer. As well, the decreased distance between the ice front and the Cochrane area would have enabled the river to carry more and coarser material to the area, resulting in the deposition of the Bighill Creek Formation. Eventually, the retreat of the glacier and subsequent increase in meltwater discharge lead to the incision of the valley fill alluvium (Stalker, 1968).

Studies by McPherson (1963), Pheasant (1968), Harris and Boydell (1972), and Boydell (1978) provided no major evidence of valley train or postglacial terrace sets in the upper Red Deer River valley. For part of the lower Red Deer

River near Empress (Figure 1-15, Site 3), though, McPherson (1968) identified two suites of non-paired alluvial terraces. The lower terrace surfaces lie 1.2 to 4.5 meters above the present floodplain. The terraces are composed of fine-grained alluvium, similar in composition to sediments forming the modern floodplain. The alluvium varies from cross-bedded medium sands to compact fine sand/clay beds. The upper terrace remnants, with treads approximately 30 meters above the present channel, are composed of poorly to moderately sorted and stratified sands and gravels, with cross-bedded structures in some locations. These deposits were thought to be of glaciofluvial origin, deposited during the ablation of the Laurentide ice sheet. Subsequent postglacial erosion was considered responsible for the terrace cutting (McPherson, 1968, p. 234).

In this area the Red Deer River valley varies greatly in width. Upstream from Empress the average valley width is approximately 1.6 kilometers, whereas downstream it widens to approximately 6 kilometers. Subsequent to the last Laurentide advance meltwaters incised the valley, which post-dates the youngest glacial drift (McPherson, 1968). The valley shape and width reflect the nature of the material into which glacial meltwater incised. Upstream the channel cut through local bedrock producing a V-shaped valley; downstream it cut through deeper glacial deposits forming a wide, extensive valley (Figure 1-18). Sands and gravels were deposited in the valley

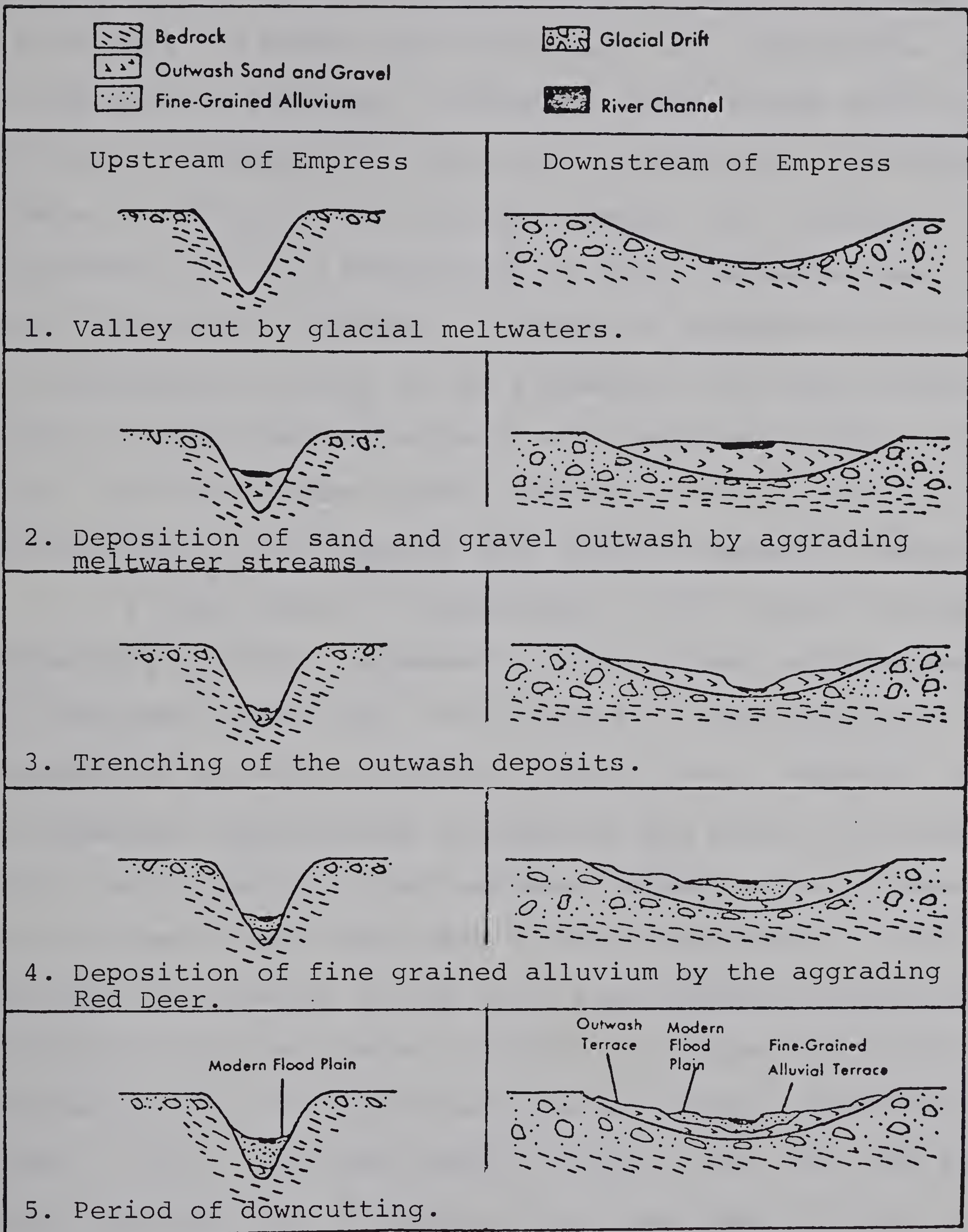


Figure 1-18. Idealized diagram illustrating the development of the valley, (after McPherson, 1968).

following the retreat of the Laurentide glacier. Trenching of the sand and gravel then took place, to a depth below that of the modern floodplain. McPherson (1968) stated that this period of degradation may have been triggered by: (1) climatic change at the end of the Wisconsin period; (2) isostatic adjustment; or (3) a combination of those two mechanisms. While the specific sequence of events is uncertain this period of degradation did lead to the formation of an upper alluvial sand and gravel terrace suite in the downstream sector of the area. In the upstream sector, because of the narrow valley, the sand and gravel deposits were almost completely removed.

A later period of aggradation in the valley followed, depositing the thick sequences of fine-grained alluvium making up the lower terrace set. The reasons for this period of aggradation are also uncertain. David (1964), suggested that a comparable aggradational sequence in the South Saskatchewan River valley resulted from isostatic rebound or an increase in the river's base level much further downstream. "As the Red Deer is tributary to the South Saskatchewan, aggradation in the Red Deer was almost certainly controlled by rate of accumulation in the South Saskatchewan Valley." (McPherson, 1968, p. 239). The most recent history of the lower Red Deer River has been that of incision below the lower terrace treads.

Alluvial terrace studies by Rains (1969) and Shelford (1975) in the Whitemud and Weed Creek valleys, respectively, both tributaries of the North Saskatchewan River, revealed

mutually consistent terrace suites which also related closely to those of the North Saskatchewan River valley near Edmonton (Figure 1-15, site 4). Westgate (1969) argued that this sector of the North Saskatchewan River valley experienced alternating phases of Late Quaternary cutting and filling and that these fluctuating periods of erosion and deposition were probably controlled by fluctuations in the position of the Laurentide ice sheet to the east. Proglacial lake ponding and river aggradation were probably favoured during periods of ice advance and still-stand, with accelerated river degradation occurring during phases of glacial retreat and subsequent lake lowering. Westgate (1969) described evidence of four river terrace elevations in the North Saskatchewan River valley at Edmonton. In a later paper (Westgate et. al., 1976) this sequence was revised, without explanation, to three main terraces. While two recent reconnaissance surveys by a number of graduate students in geography, University of Alberta, confirmed the existence of at least three terrace units more detailed work will be required to confidently exclude the fourth, identified earlier by Westgate (1969). A major related problem is that long-continued channel migration of the North Saskatchewan River (Thomson and Townsend, 1978) has promoted very efficient destruction of the older alluvial terrace remnants in that valley. This suggests that the relatively underfit tributary creeks are more likely to have complete terrace sequences preserved in their valleys (Rains, pers. comm., 1978).

In the tributary Whitemud Creek valley (Figure 1-19) numerous remnants of three paired alluvial terraces are very well preserved. Two remnants of a higher, older terrace were also identifed by Rains (1969). The lowest, youngest, terrace grades downstream to merge with the lowest North Saskatchewan River terrace described by Westgate (1969) and Westgate, et. al., (1976). A downstream projection of the three higher Whitemud Creek terrace elevations revealed a mutual divergence in that direction and their probable linkages with the three highest North Saskatchewan River terraces (Rains, 1969, p. 203; Westgate, 1969).

Exposures of alluvium making up the four discrete terrace units of Whitemud Creek valley typically include a basal lag gravel layer above a sharp contact with either Cretaceous bedrock or Pleistocene tills. Alluvium above the gravels consists of sands, silts and clays previously deposited in point bar and overbank floodplain environments. Bedding structures are not well developed in the fine-grained alluvium except for occasional sub-units with good, sub-horizontal laminations. The alluvium of each terrace unit characteristically maintains total thicknesses of 1.5 to 3 meters. Grain-size analyses of the fine-grained alluvium of the main terrace units showed that the lowest terrace deposits are slightly richer in sand than the higher deposits (Rains, 1969).

The distribution of terrace remnants in this valley led Rains (1969) to conclude that at least three and probably

four, distinct episodes of aggradation and/or channel stability have occurred in late Quaternary time. He outlined a tentative chronology of this valley's evolution which remains to be confirmed or modified by further studies, hopefully incorporating radiocarbon dates (Rains, pers. comm., 1978).

In his study of the Weed Creek valley (Figure 1-19), Shelford (1975) identified four paired alluvial terraces which mirrored closely the Whitemud Creek terrace suite. In particular this valley has very extensive preservation of the uppermost terrace, considered to be correlative of the two, uppermost terrace remnants in the Whitemud Creek valley. The character of alluvium also closely parallels that of Whitemud Creek, and there can be no doubt that the two valleys have evolved in concert—in response to controls which must have been of regional rather than local importance.

The terraces of Whitemud Creek, Weed Creek, and the North Saskatchewan River near Edmonton are all very closely related, as indicated by the similar geomorphic chronologies of the two tributary basins. Alternating periods of aggradation and degradation in the Whitemud and Weed Creek basins must have been at least partly controlled by local base level fluctuations of the North Saskatchewan River to which they are adjusted.

Two paired, late Pleistocene, valley train terraces in the Athabasca River valley near Hinton (Figure 1-15, Site 5) were described by Stene (1966) and Roed (1968). The upper

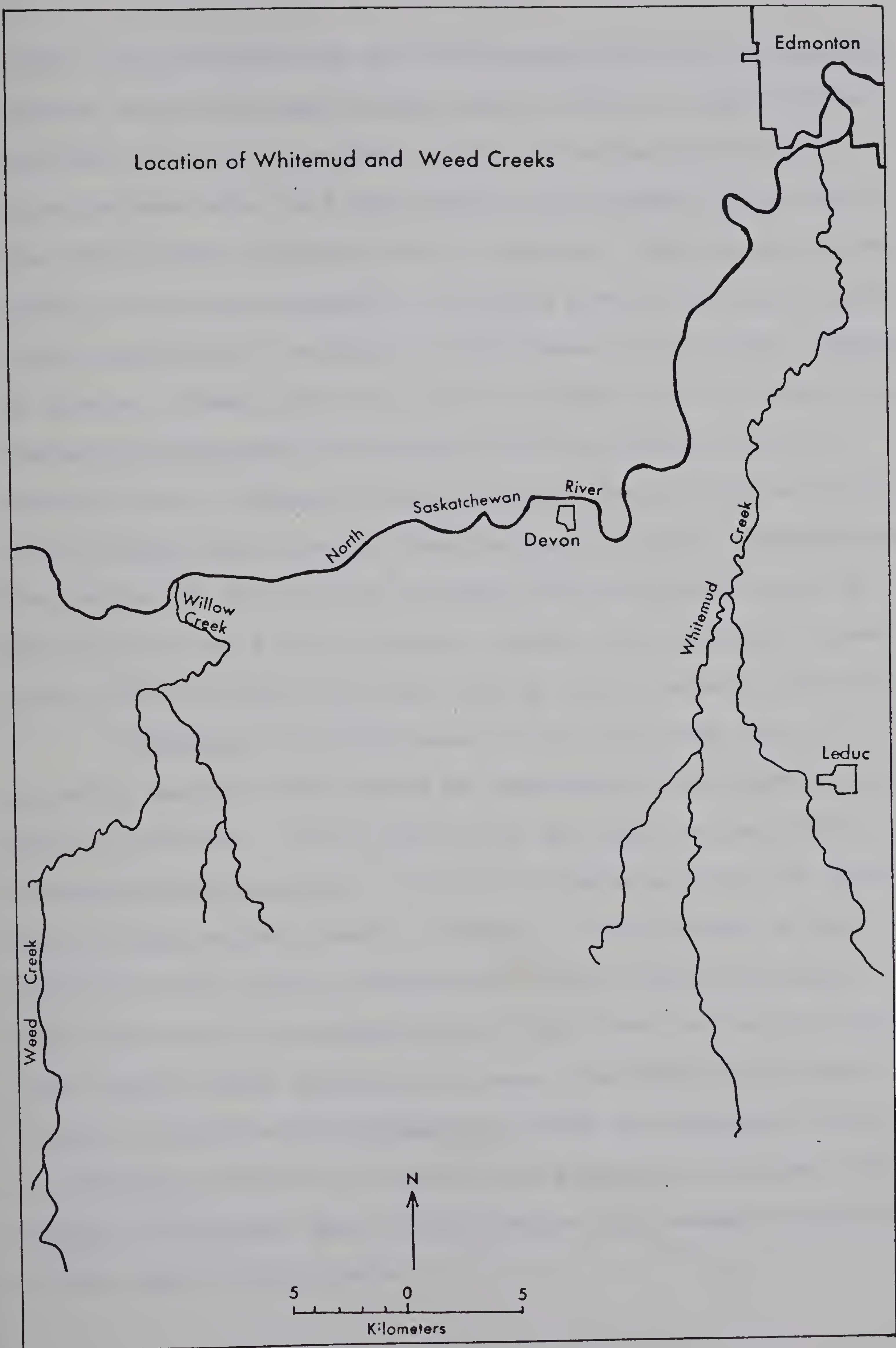

Edmonton
Location of Whitemud and Weed Creeks
North Saskatchewan River
Devon
Willow Creek
Whitemud Creek
Leduc
Weed Creek
N
5 0 5
Kilometers

Figure 1-19

valley train terrace may be traced upstream to an end moraine complex while the lower valley train surface extends further upstream to an indeterminate origin. The deposits of the upper terrace were laid down during a prolonged still-stand of the Cordilleran, Athabasca Valley glacier. The terrace gravels appear to have been deposited during a period of rapid aggradation and display the short, discontinuous cross-strata typical of glacial outwash deposits. A thin upper unit of coarse, lag, channel gravels overlies the bulk of the outwash deposits (Stene, 1966). Climatic change brought about continued retreat of the Cordilleran glacier from its earlier still-stand position. The retreat of the glacier provided excessive quantities of meltwater to the fluvial regime, eroding the previous outwash valley fill to leave the upper valley train terrace remnants.

Incision of the Athabasca River continued until a probable, second, still-stand or readvance of the Cordilleran glacier occurred. During this time the lower valley train sediments were deposited. This unit consists mainly of poorly sorted cross-bedded gravels. Again, a later change of the Athabasca River regime accompanied further glacier wastage. This resulted in a rejuvenation of the river, entrenching the lower valley train sediments to leave the lower valley train terrace remnants which extensively flank the Athabasca River in the Hinton district. Lower, less extensive terraces, presumably of Holocene ages, occur between the present floodplain and the lower valley train.

Rutter (1977) described two major terraces of the Peace River in northeastern British Columbia (Figure 1-15, Site 6). The upper glaciofluvial terrace was considered to have been deposited during a retreat phase of the last major Cordilleran glaciation of the area. The terrace deposits consist of interbedded sands and gravels, typical of braided channel deposition. The lower terrace unit, 4.5 meters above the present floodplain, is relatively continuous along the Peace River. Rutter (1977) speculated that these deposits may have formed during a limited period of glacial advance and that subsequent downcutting to produce the terrace occurred during glacial retreat, or in response to isostatic adjustment of the Cordilleran region.

1.3.3 General Conclusions

In general there is considerable agreement among previous studies that late Quaternary alluvial terrace development in Alberta resulted directly from the influence of Cordilleran and Laurentide glaciers on major catchments. Stene (1966), Stalker (1968), Rutter (1972) and Alley and Harris (1974) all concluded that terrace development in those valleys bordering the foothills area of Alberta resulted from fluctuations in channel sediment/discharge conditions induced by fluctuations in the wastage of related Cordilleran glaciers. The periodic damming of rivers by receding Laurentide ice lobes, coupled with the fluctuations of sediment/discharge associated

with Cordilleran glacial retreat, were cited as the probable cause of terrace development by Rutter (1977) and Alley and Harris (1974). Conversely, Stene (1966) and Stalker (1968) argued that alluvial terrace development, in their respective valleys, resulted during a period of glacial advance and subsequent still-stand phase; ice damming by a Laurentide ice lobe exerting no significant influence on terrace development. In view of these conflicting interpretations of Cordilleran ice movements leading to terracing, then, it may only be concluded that there exists a need for a better understanding of Cordilleran glacier movements along the major valley systems concerned, and corroborative time correlations of these movements.

There seems to be little argument that the Laurentide glacier acted as a base level control, directly influencing terrace development along the major river basins to the east. The controversy which exists regarding terraces of the North Saskatchewan River is not raised over the movements of the Laurentide glacier but on the number of paired terrace sets identified. Initially Westgate (1969) identified four terrace sets, but later changed this interpretation, suggesting instead that only three terraces exist. Contradictory to this, work by Rains (1969) and Shelford (1975) on nearby tributary streams, which would have been directly controlled by the level of the North Saskatchewan River, show clear evidence of four main terraces.

As was suggested earlier the interpretation of alluvial terrace development in Alberta during the late Quaternary period is partly in a state of disarray. A variety of authors, working in various sectors of the province, have presented their findings without attempting to correlate these results with other regions. All too often the data presented have been tenuous, preventing critical evaluation and allowing only speculative suggestions for alternative interpretations. As well, with the exception of the North Saskatchewan River tributaries studies near Edmonton, interpretations have been based largely on evidence from portions of major valley systems. Complementary work on terrace formations along tributary systems, which are directly linked to the variations of the main channels, have been largely ignored.

CHAPTER II

RESEARCH METHODS, HYDROLOGY AND BEDROCK GEOLOGY

2.1 Introduction

The Athabasca River originates in the Front Ranges of the Rocky Mountains and flows 1450 kilometers to the north-east, entering Lake Athabasca near the Alberta/Saskatchewan border (Figure 2-1). This study examines an approximately 200 kilometer long sector of the river valley from Hinton to Whitecourt (Figure 2-2). Within this sector the river channel pattern progresses downstream from a relatively single-channel form, with very occasional islands, to a quasi-braided type with numerous islands, spool bars and point bar complexes (Plate 2-1). The river valley is presently being entrenched in the upstream sector of the reach, but is not obviously degrading or aggrading in the lower valley reach (Bray, 1972). The channel slope within the study sector between Hinton and Obed Ferry is somewhat gentler than from Obed Ferry to White-court. Figure 2-3 shows a generalized water surface long profile at low stage in relation to the valley tops. The river is contained within a generally steep-sided valley, varying from 5 to 9 kilometers in width. Figure 2-4 depicts selected, representative, valley cross-profiles.

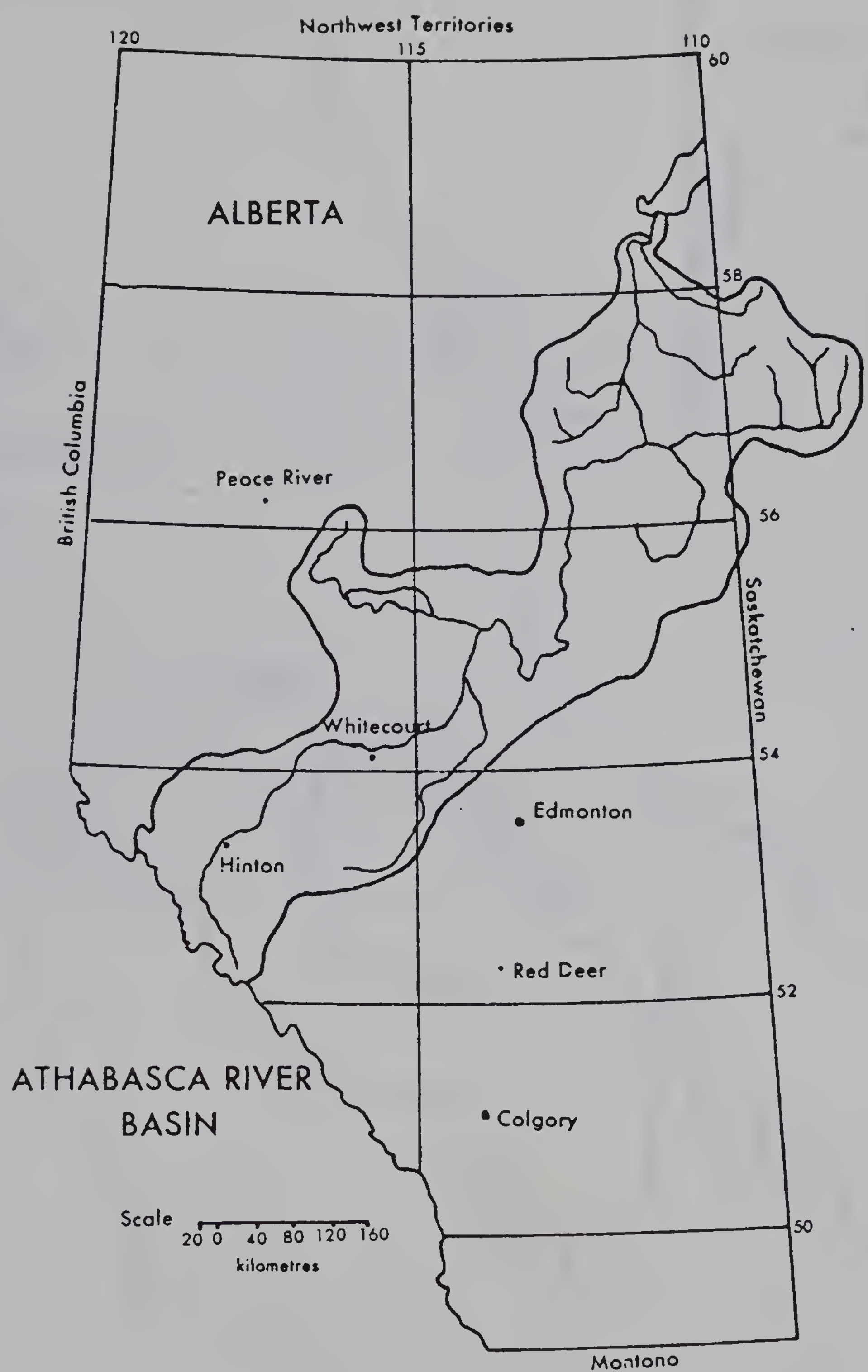


Figure 2-1, (after Hillman et. al., 1978).

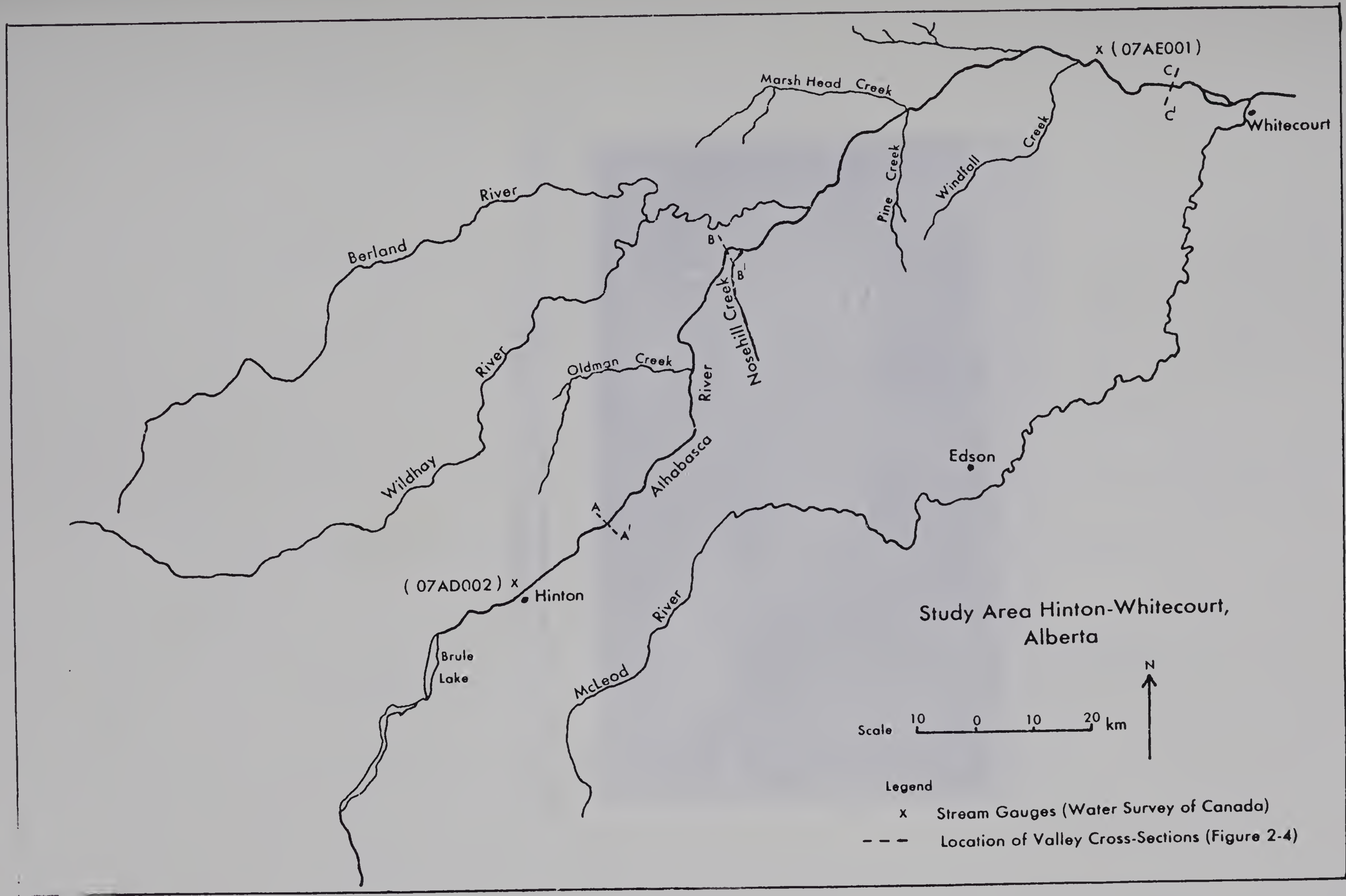
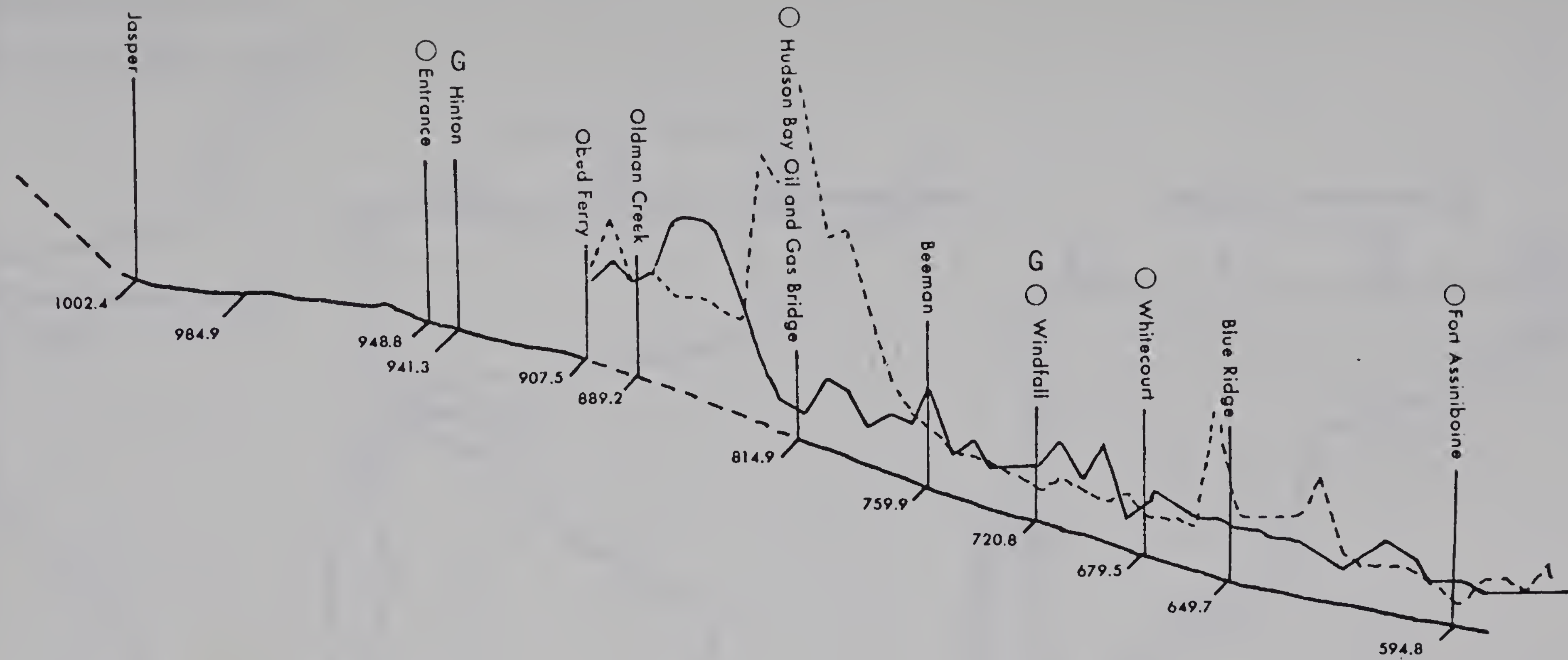

Study Area Hinton-Whitecourt, Alberta
x (07AE001)
Whitecourt
Marsh Head Creek
Pine Creek
Windfall Creek
Nosehill Creek
Berland River
Wildhay River
Oldman Creek
Athabasca River
McLeod River
Edson
(07AD002) x
Hinton
Brule Lake
A
A'
B
B'
C
C'
N
Scale 10 0 10 20 km
Legend
x Stream Gauges (Water Survey of Canada)
- - - Location of Valley Cross-Sections (Figure 2-4)

Figure 2-2

Plate 2-1. Quasi-braided channel with spool bar and point bar complex.



Long profile of the Athabasca River, Jasper - Fort Assiniboine

LEGEND

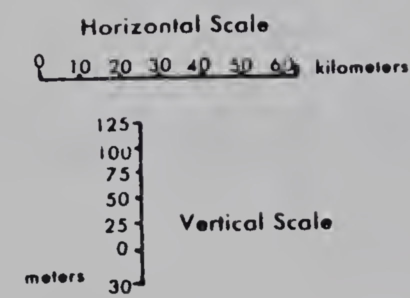
Bridge...................................○
Gauging Station.........................G

Water Surface Elevation............

Water Surface Elevation, Estimated.....

Valley Top-Left Side..................

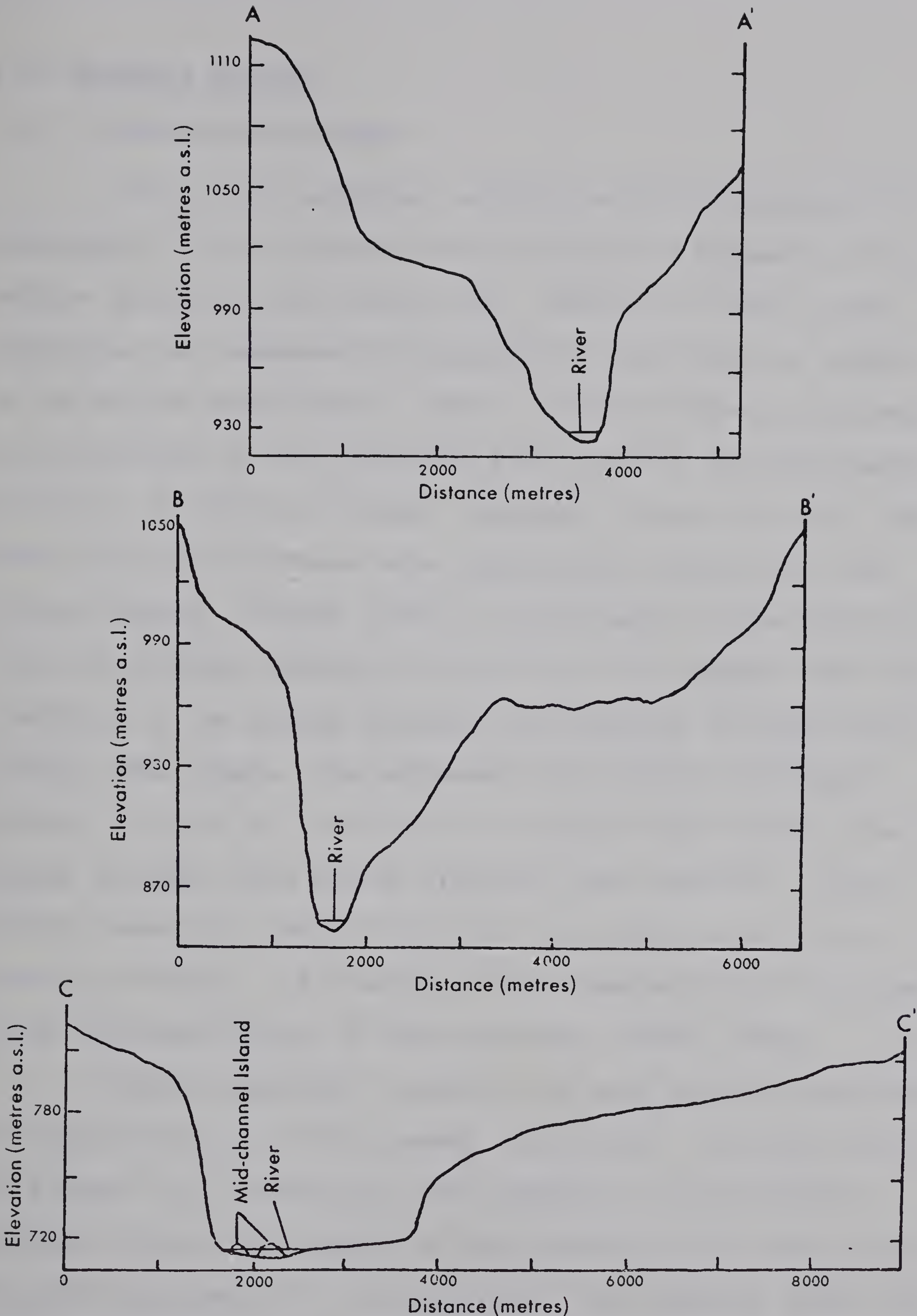
Valley Top-Right Side.................



V.E.=250X

(after Kellerhals, et. al., 1972).

Figure 2-3



V.E.=22X

Athabasca River Representative Valley Cross-Profiles

Figure 2-4

2.2 Research Methods

2.2.1 Problem Formulation

Much of the existing research on the Quaternary geomorphology of the Athabasca River valley has focused on the western portion of the study area. Hector, in 1859, first recognized and assessed the heights of three terrace levels in the Hinton area (Stene, 1966). In 1898, McEvoy confirmed the glaciation of the Athabasca River valley, by Cordilleran glaciers, in the Hinton area. However, it was not until 1960 that the Hinton terraces were specifically correlated with glacial events. Taylor (1960), in his paper on the Pleistocene lakes of northern Alberta, was the first to suggest that the elevation of the Hinton terraces corresponded to former proglacial lake levels. He suggested that during two stable periods, of what he referred to as Glacial Lake Miette, the Hinton terraces were carved along the lake margins. Later, though, Mountjoy (1964), discounted the terraces as being glacially formed, and considered the Athabasca River terraces to be modified moraine or kame terraces (Stene, 1966).

Little geomorphic research has been done on the remaining major portion of the present study area. St-Onge (1972), in a paper on a proglacial lake sequence of north-central Alberta, traced the retreat of the Laurentide ice sheet, and related development of the proglacial lake systems, along part of the Athabasca River valley. He did not attempt to correlate river terrace elevations with known proglacial lake levels

downstream.

Throughout the Athabasca River valley terrace remnants are primarily the product of interactions between late Quaternary, Cordilleran, glaciofluvial activity in the headwater parts of the valley and proglacial, lacustrine, base level controls associated with the Laurentide ice-sheet downstream. Stene (1966) examined river terrace development relative to the advance and retreat of the Cordilleran, Obed glacier between Entrance and Hinton (Roed, 1975). Extensive glaciofluvial materials deposited in that section of the valley illustrated that meltwater drainage was not obstructed by Laurentide ice in close proximity downstream. Stene (1966) suggested that immediately east of Hinton a change in one or more hydraulic factors resulted in the terracing of the lower valley. He believed that climatic change led to the production of increased meltwater flow from the retreating Obed glacier and this promoted valley fill trenching.

Another possible explanation for the development of these terraces may be changes of local base levels downstream. The northeasterly trending slope of the land, and the northwest-southeast alignment of Laurentide marginal moraines (Bretz, 1943), suggest that the Laurentide ice wasted in a northeasterly direction away from the Rocky Mountains. As the ice front retreated to the northeast, exposing surfaces at lower and lower elevations, proglacial lakes formed temporary base levels for the east-flowing rivers. With the lowering of

a proglacial lake, rivers such as the Athabasca would erode and create terraces. The depth of river degradation would have been limited by the level of the next proglacial lake into which it flowed. During extended periods of proglacial lake ponding river aggradational processes would be favoured, leading to the development of alluvial fills in the valleys upstream of lake margins. Therefore, it is suggested that temporarily stable, proglacial, lake levels probably acted as the main controls of the downstream extent of alluvial valley fills (Figure 2-5).

Thus the major objectives of this study were to map the distribution of alluvial terrace remnants between Hinton and Whitecourt and to interpret their probable relationship to major proximal and distal controlling factors. Figure 2-6 outlines a variety of predicted, possible terrace surface relationships which might result from varied upstream and downstream controls. The observed field relationships of the Athabasca River terraces are detailed in Chapter IV.

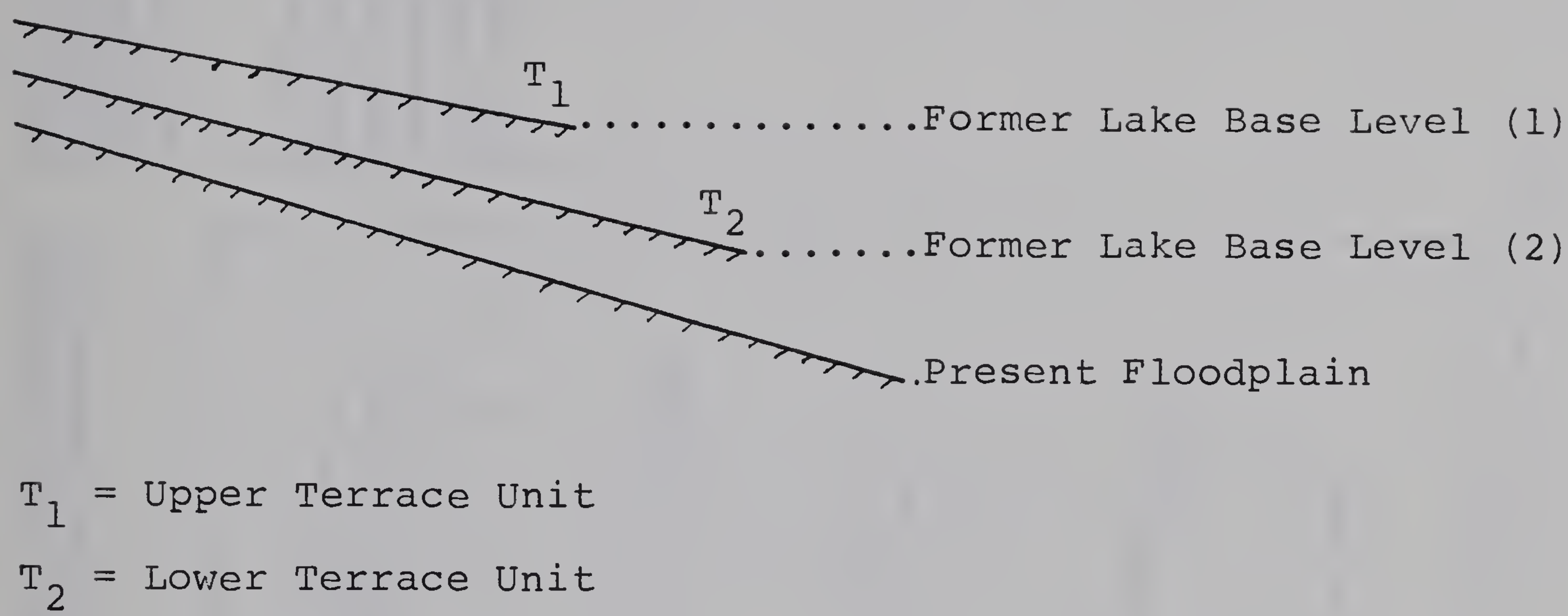


Figure 2-5. Hypothetical relationship between remnant terrace units and former, temporary, stable, proglacial, lake levels.

2.2.2 Field Methods

The initial phase of the project involved the production of base maps from aerial photographs (scale 1:31680) and topographic maps (NTS 83F, 83J and 83K). From aerial photographs, the interpretation and distribution of probable terrace

Upstream Controls:
Cordilleran ice advances and retreats,(Stene, 1966; Roed 1968,1975).

Stene (1966) outlines two retreat phases of the Obed glacier leading to the development of an upper and lower paired terrace sequence.

Roed's (1968,1975) Drystone Creek advance was short lived, with outwash terrace development during the retreat phase. There is, however, no indication of end moraines being linked with Stene's Obed glaciation. There are no dates available for the times of glacial retreat and advance as outlined by Stene (1966) and Roed (1968, 1975).

Downstream Controls:
Laurentide ice marginal positions and proglacial lake extent,(St-Onge,1972).

St-Onge (1972) determined the ice marginal positions of the Laurentide ice sheet from deltaic and spillway outlet elevations.

Hypothetical terrace relationships in the Athabasca Valley tested by field mapping

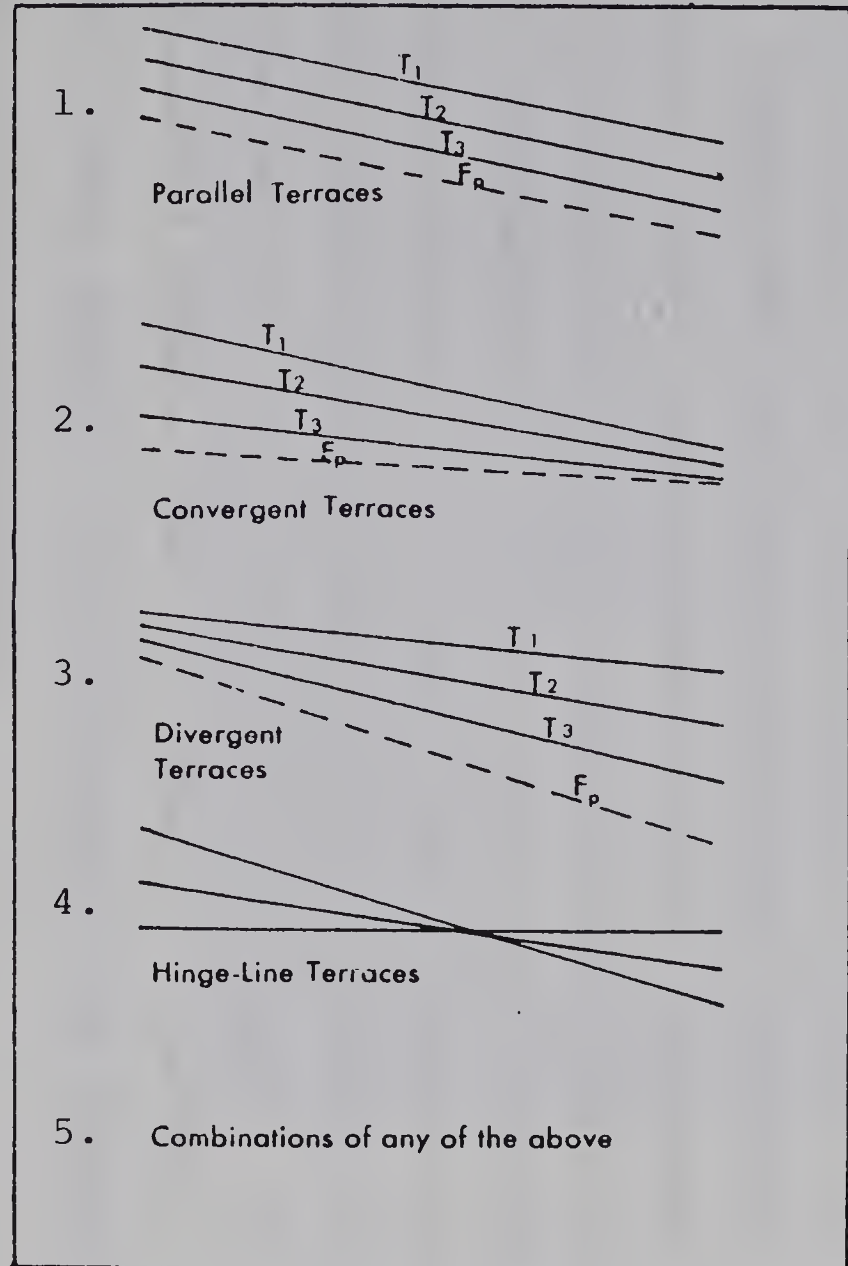


Figure 2-6. Possible terrace relationships along the Athabasca River.

remnants, lacustrine forms, deltas, spillway channels, moraines and ice-contact features in the immediate vicinity of the valley were recorded on the base maps. To complement base map information longitudinal valley-top and river profiles were employed from reports by Kellerhals et. al., (1972) and Neill (1973).

The field work was carried out from May through to August, 1978. The bulk of the field program entailed the location, logging and photographing of alluvial exposures, demarcation of terrace tread remnants on aerial photograph overlays and surveying of valley cross-sections, with a Paulin altimeter, to determine the relative elevations of the floodplain, terrace treads and valley tops. The altimetric surveys were not of a high level of accuracy because logistical difficulties prevented the use of a base station barograph. However, because the terrace treads are generally separated vertically by many meters, the inaccuracies of the uncontrolled altimetric surveys are not critical.

A major portion of the lands bordering the Athabasca River valley in the study area are under long term lease to the North Western Pulp and Power Company Ltd., of Hinton and the Hudson Bay Oil and Gas, Chevron and Petrofina Corporations, of Calgary (Mr. G. Firth, Chevron field foreman, pers. comm., 1978). Extensive logging operations in the vicinity of Hinton, and the potential threat of poisonous gas leaks (Plate 2-2) throughout much of the area, have kept land development

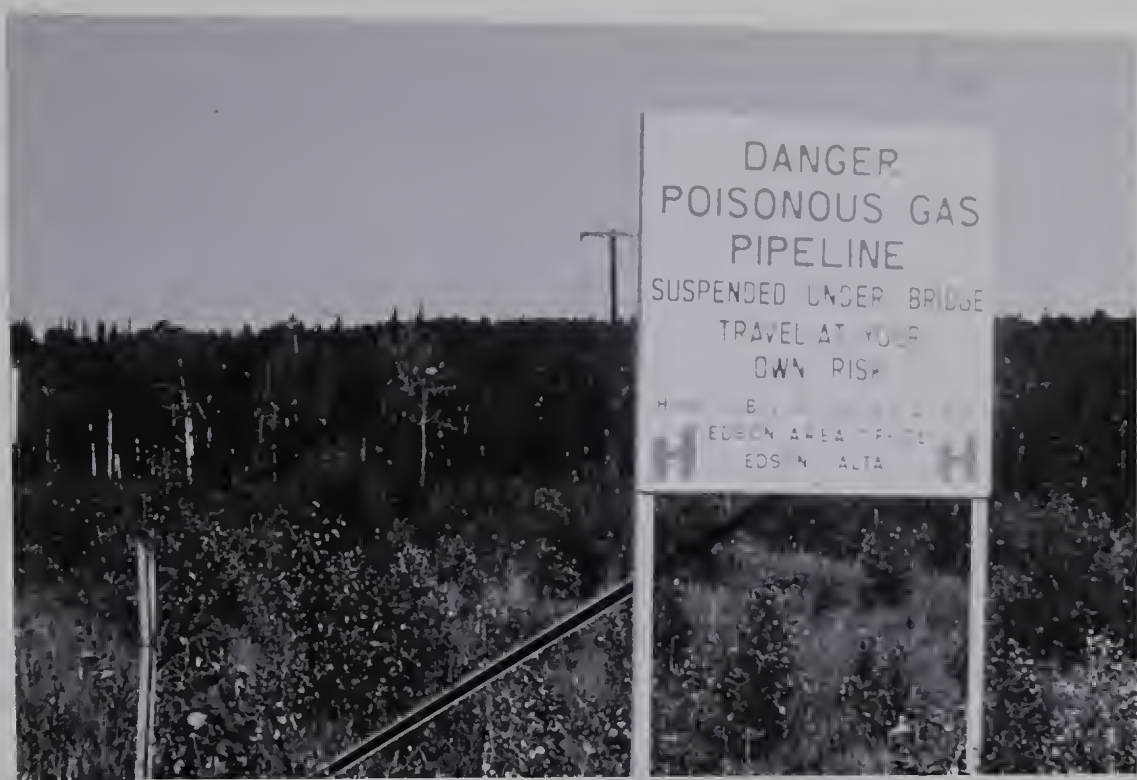


Plate 2-2. Identification marker of potential hazard encountered in the study area.

including agriculture to a minimum. Consequently, much of the area remains in its natural state. The terrain is rough and difficult to traverse and road access is very limited. Because of the limited access several modes of transportation were employed during the field season. Initial reconnaissance of the study reach was undertaken using a jet boat. The principal portion of the study was carried out by travelling downstream in a canoe, establishing a series of temporary tent camps, and hiking the major forest trails and seismic cut lines. Later, all river access roads and bordering highways were traversed by car.

2.3 Hydrology

The Water Survey of Canada (WSC) maintains two stream gauges within the study sector of the Athabasca River. One gauge (WSC07AE001) is situated near Windfall Creek, 26 kilometers upstream of Whitecourt. The second gauge (WSC 07AD002) is located at Hinton (Figure 2-2). Major tributaries to the Athabasca River in the area are the relatively shallow and fast flowing Berland and Wildhay Rivers, and the Oldman, Pine and Windfall Creeks. During the winter months the rivers and creeks are completely covered by ice. While a small base flow is usually maintained by the main rivers during winter months, it is not until mid-April, as the snow begins to melt, that flows begin to increase. Consequently, hydrographs included here will detail representative discharges of only summer months.

Figure 2-7 shows the 1974 hydrographs for the Athabasca River and Pine Creek. The shape of the hydrograph for Pine Creek is typical of small streams in the area (Hillman et. al., 1978). The hydrograph shows that the stream's most sustained high discharges are dominated by spring snowmelt inputs with additional, shorter duration, high flows accompanying summer storms. During the snowmelt period flows from the many small tributary streams combine to contribute to the characteristic shape of the May hydrograph for the Athabasca River. By late May and early June most of the snow is gone from the study area and the hydrograph for Pine Creek shows that, except for occasional rainstorms, recession is taking place. The hydrograph for the Athabasca River, however, shows that its peak flows do not occur until mid-June. The later onset of warmer temperatures at higher altitudes in the Rocky Mountains leads to increased snow and glacier melt inputs in June. Combined with spring rain-storms this determines the gradual decline of river discharges over the summer months. Occasionally large rainstorms, such as the one which occurred in August, 1969, contribute sufficient runoff to the Athabasca River to equal or exceed that normally produced by snow and glacier melt (Figure 2-8).

Pertinent hydrological data including mean discharge, minimum flows and maximum flows are summarized in Table 2-1. Table 2-1 reflects the water-yielding characteristics of the upstream areas. The drainage area upstream of Hinton is one

TABLE 2-1

Annual Flow Data For Athabasca and McLeod Rivers, 1961-1976

(from data supplied by Water Survey of Canada)

	Athabasca River					
	Windfall Gauge 07AE001 Drainage area: 18 100 km^2			Hinton Gauge 07AD002 10400 km^2		
Year	Annual flow ($x10^9$ m^3)	Max. flow (m^3/S)	Min. flow (m^3/S)	Annual flow ($x10^9$ m^3)	Max. flow (m^3/S)	Min. flow (m^3/S)
1961	7.40	949	36.0	5.28	804	19.7
1962	8.10	966	32.8	5.16	765	11.0
1963	7.45	917	29.7	5.48	719	10.8
1964	8.35	1230	23.4	5.53	844	15.0
1965	11.26	2130	30.3	6.70	900	15.7
1966	9.60	1190	38.5	6.01	765	17.2
1967	7.83	1170	28.2	6.23	855	24.4
1968	7.48	1180	29.4	6.03	1000	27.5
1969	7.08	1760	31.1	5.56	997	31.1
1970	5.85	946	34.5	4.45	776	19.8
1971	8.97	1700	32.6	5.86	895	21.5
1972	8.82	1870	37.4	6.14	1200	27.8
1973	7.18	1170	51.0	4.85	963	29.7
1974	7.97	1250	42.5	5.85	1070	26.9
1975	5.92	850	39.6	4.60	736	28.3
1976	7.92	1042	42.2	5.60	640	21.0
Mean	7.95	1270	35.0	5.58	871	21.7
M.A.R.[1]	439 mm			537 mm		

[1] Mean annual runoff

Source: Hillman, et. al. (1978)

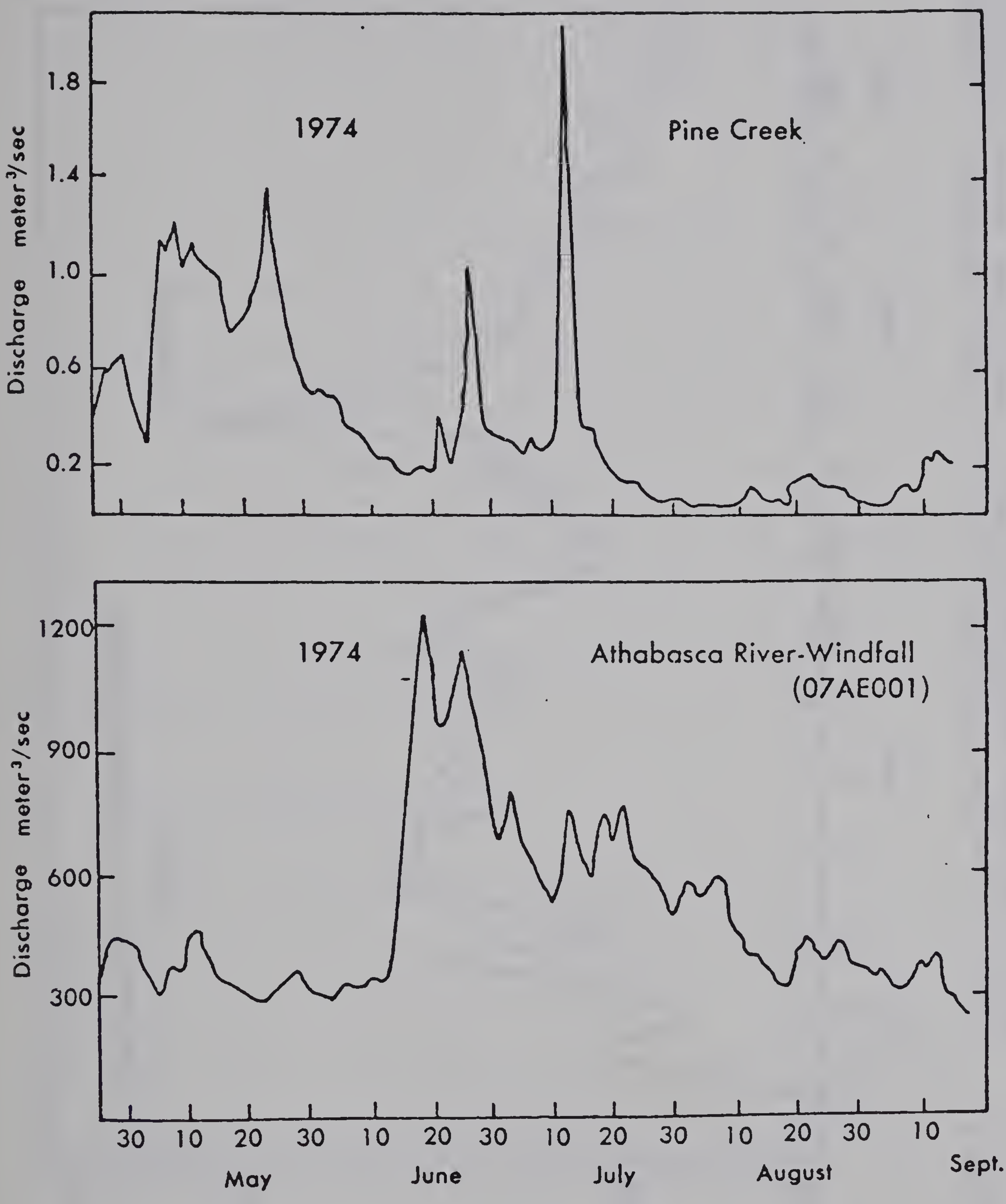


Figure 2-7. Streamflow data for the Athabasca River and Pine Creek, 1974 (includes data supplied by Water Survey of Canada), (after Hillman et.al., 1978).

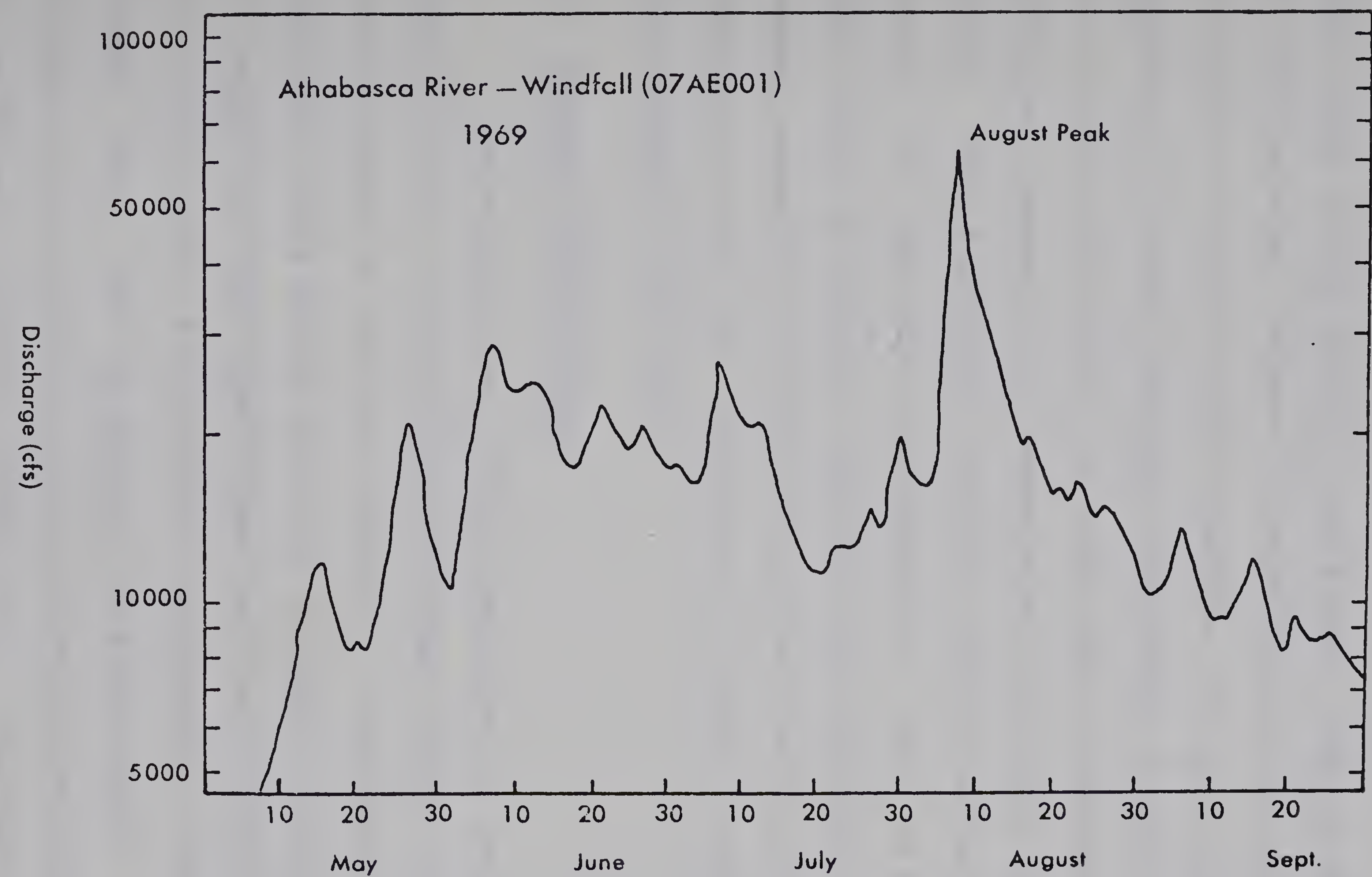


Figure 2-8. Effect of summer precipitation on river discharge, (adapted from hydrograph supplied by Water Resources Branch, Ministry of the Environment, Canada).
(1 cfs. = 0.028 m^3/sec)

of the highest water-yielding areas in Alberta (Hillman, et. al., 1978). However, the increased flow past the Windfall Creek gauge is not greatly significant, despite the increase in drainage area, because of the moderate water-yielding nature of the area between the two stations. An approximation of the proportion of total discharge contributed to the Athabasca River system from within the study area is indicated by the difference between flows measured at the Hinton and Windfall gauges. Data from these stations, for the period April to August, 1977, are shown in Figure 2-9. Plots of mean monthly flows (March to October, 1961-1974) for the Hinton and Windfall gauges, presented in Figure 2-10, show that the flow regime described for the 1977 hydrograph is similar to the average flow regime for 1961-1974.

2.4 Bedrock and Structure

The western part of the study area is located in the Foothills Belt, a structurally deformed region lying along the eastern border of the Rocky Mountains. Sub-parallel thrust faults and folds strike generally northwest-southeast. Normally the anticlinal folds form ridges and the synclinal folds valleys, but to some extent the folds are independent of the topography (Irish, 1965). West of Hinton, the Pedley Fault, a thrust fault dipping to the southwest, occurs along the northeastern limb of the Entrance Syncline (Figure 2-11). The fault extends from the southeast to the Athabasca River but has not been traced

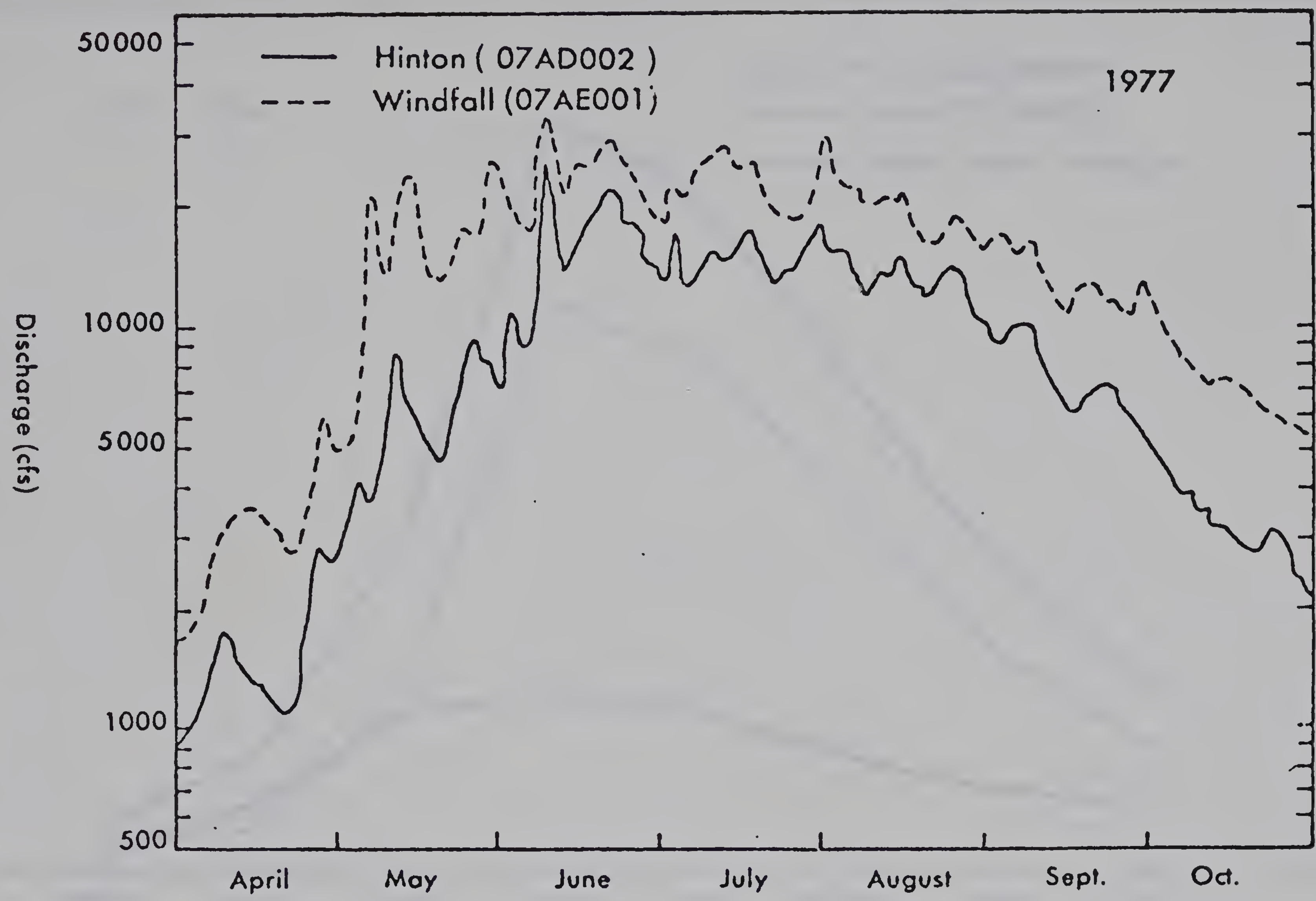


Figure 2-9. Differences in discharge of the Athabasca River at Hinton and Windfall, 1977, (adapted from hydrographs supplied by Water Resources Branch, Ministry of the Environment).
(1 cfs. = 0.028 m^3/sec)

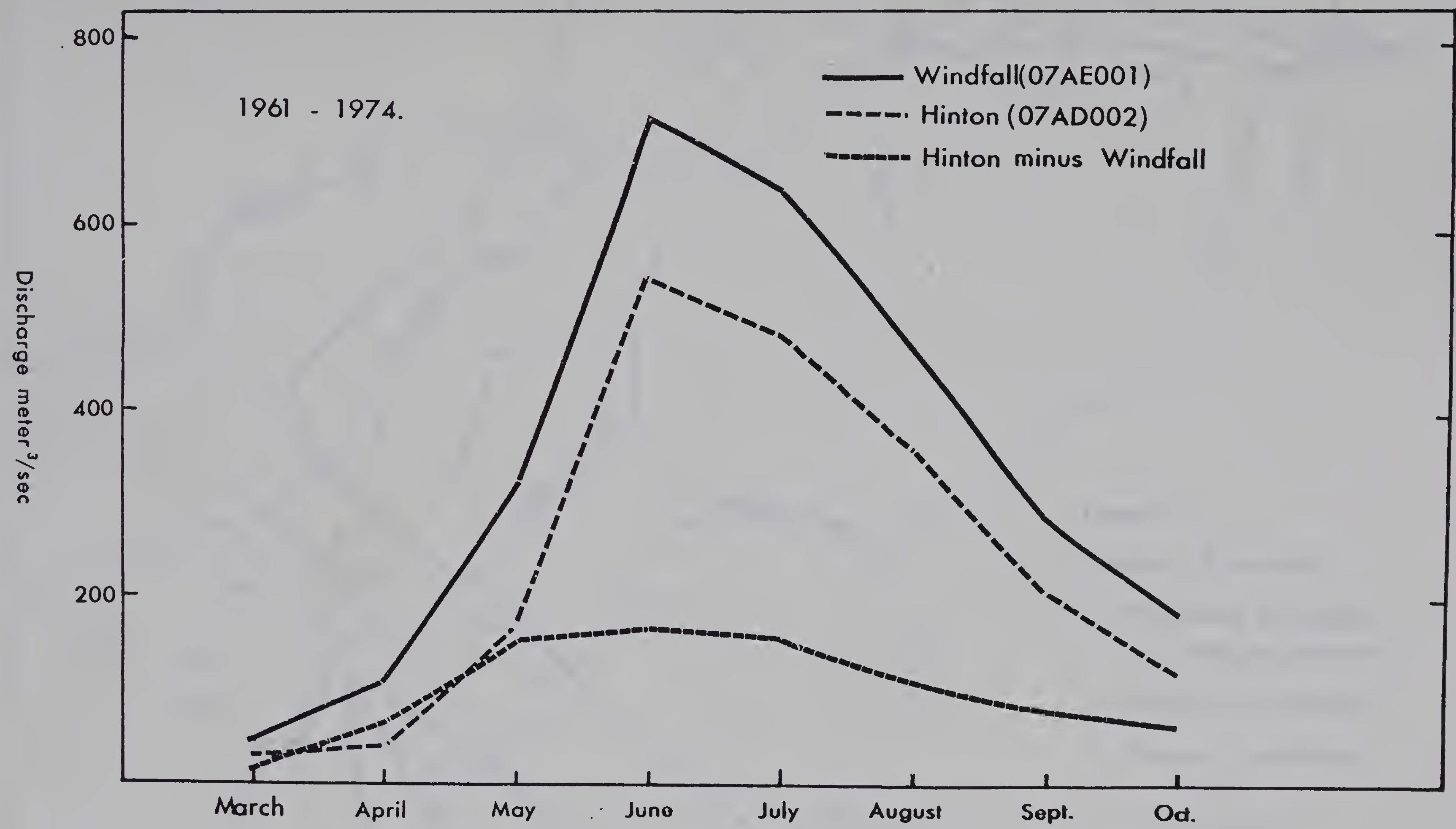


Figure 2-10. Hydrographs for the Athabasca River (1961-1974) showing the differences in discharge at Windfall and Hinton, (after Hillman et. al., 1978).

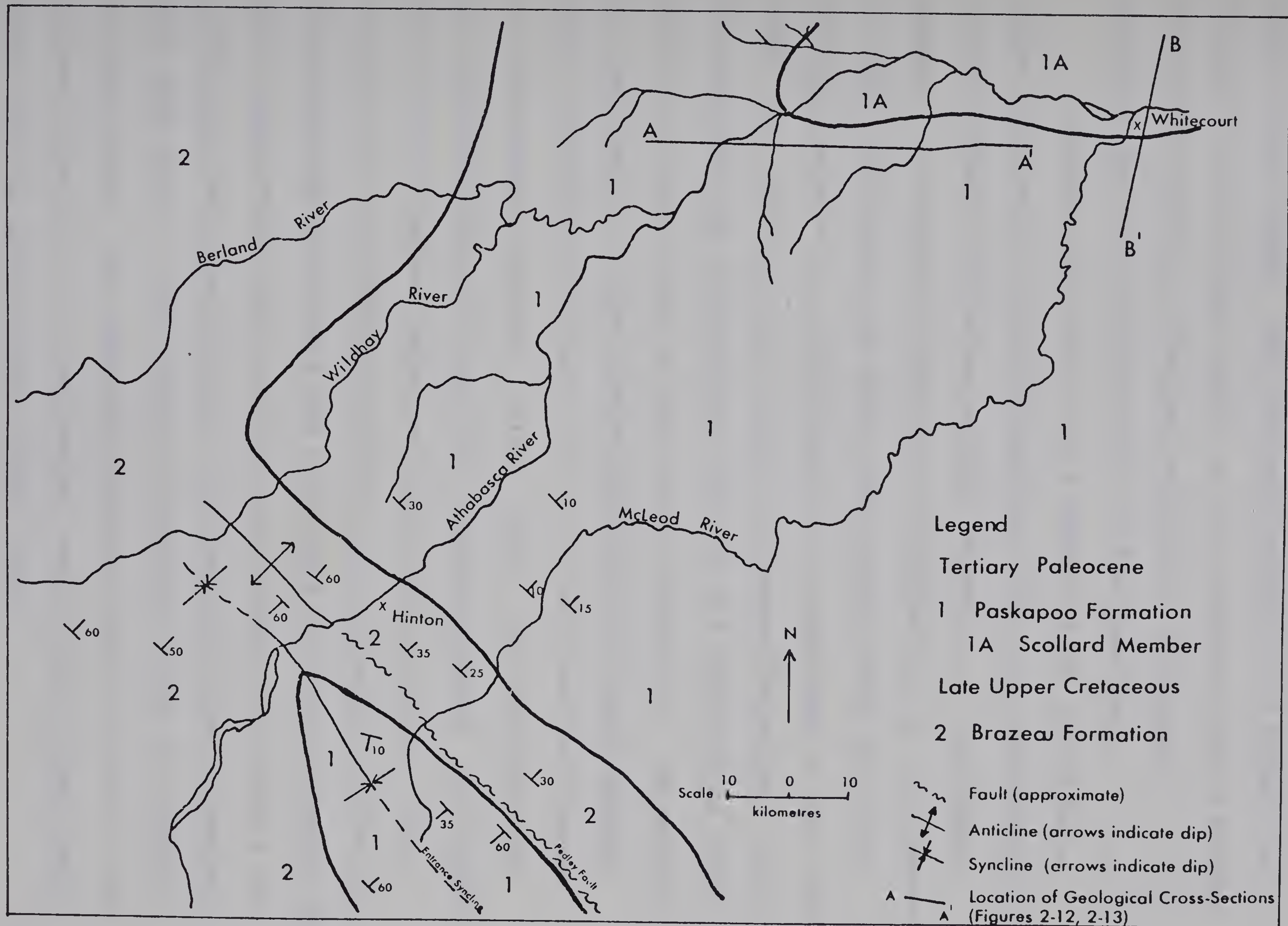


Figure 2-11. Bedrock Geology, (after Irish, 1965; Tokarsky, 1977a, 1977b).

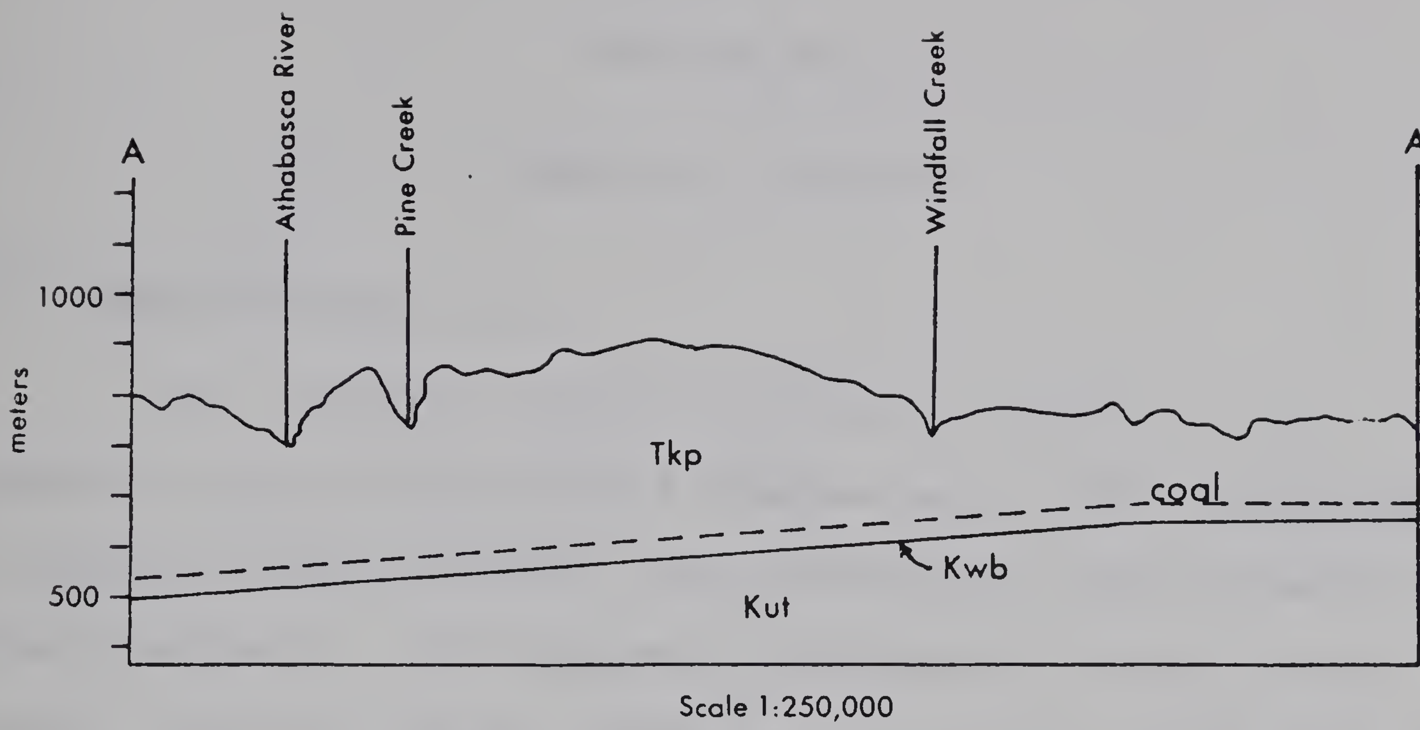
north of the river. The Entrance Syncline is a broad open syncline which plunges gently to the southeast and is bounded on the northeast by a faulted anticline. Strata of Tertiary age occur in the trough of the syncline.

Bedrock for most of this western area is the Late Upper Cretaceous Brazeau Formation, which lies conformably over the Upper Cretaceous, Wapiabi Formation. The Chungo Member of the Wapiabi Formation is overlain by 20 to 30 meters of soft, dark green and grey, sandy shales which are transitional between the underlying marine sandstones of the Chungo Member and the overlying non-marine Brazeau Formation. Directly overlying the transitional beds are the distinctive pebble beds and conglomerates of the Brazeau Formation. The amount of conglomerate in the Brazeau Formation decreases upwards until sandstones predominate with only thin beds and lenses of conglomerate included (Irish, 1965). The conglomerates are composed mainly of chert and quarzite pebbles, the individual beds and lenses having no great lateral extent. The remainder of the formation consists of interbedded medium to coarse-grained sandstones and grey shale with minor amounts of black carbonaceous shale and thin coal seams. The shales are mainly grey and greenish grey, but occasional black carbonaceous beds occur. Thin, impure, ironstone layers are present in some places. The uppermost stratum classed as part of the Brazeau Formation is a distinctive, massive bed of grey sandstone about 20 meters thick (Irish, 1965).

To the east of Hinton (Figure 2-11) the Brazeau Formation is in turn overlain by the Paskapoo Formation of Paleocene, Tertiary age (Figure 2-12). This formation consists of freshwater, grey to greenish grey, thick, bedded, calcareous, cherty sandstones; grey and green siltstones and mudstones; with additional minor conglomerates, thin limestones, coal and tuff beds. In places a disconformity has been observed at the base of the Paskapoo Formation, but where neither this disconformity nor diagnostic fossils are found it may be difficult to separate the Brazeau and Paskapoo Formations.

In the Whitecourt area, the Scollard Member of the Paskapoo Formation, overlies the Brazeau Formation (Figure 2-13). The freshwater strata consist of grey, feldspathic sandstones and dark grey, bentonitic mudstones with thick coal beds (Tokarsky, 1977b).

Unconsolidated sediments, mainly of glacial, glaciofluvial and glaciolacustrine origins, mantle much of the bedrock in the study area. The gravels within these materials consist mainly of quartzites, conglomerates, sandstones, limestones and cherts, all of which are characteristic of the uppermost bedrock formations. To the east, igneous and metamorphic clasts of Canadian Shield provenances form a minor proportion of the surface deposits. These surficial deposits are described in detail in Chapter III.

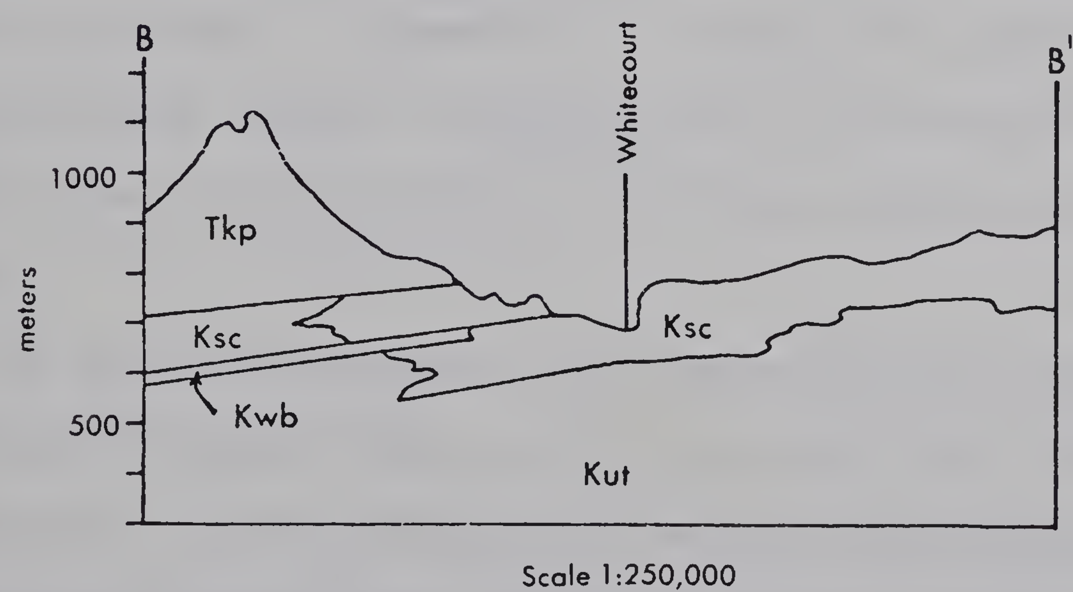


Tkp Paskapoo Formation

Kwb Battle Formation

Kut Brazeau Formation

Figure 2-12. Geological cross-section along line AA', (after Tokarsky, 1977a).



Tkp Paskapoo Formation

Ksc Scollard Member

Kwb Battle Formation

Kut Brazeau Formation

Figure 2-13. Geological cross-section along line BB', (after Tokarsky, 1977b).

CHAPTER III

SURFICIAL GEOLOGY

3.1 Introduction

The surficial deposits within and bordering the Athabasca River valley form a complex and diverse assemblage of glacial, glaciofluvial, glaciolacustrine and nonglacial, recent sediments. The following discussion of the glacial history and stratigraphic units of the general study area is based on the tentative time-stratigraphic correlation of Table 3-1. The stratigraphic section and plate localities, selected for illustrative purposes in the text, are noted in Figure 3-1. Details of the stratigraphies of all sections examined by the author are included in Appendix A. Figure 3-1 shows the location of areas mapped in Figures 3-2 to 3-7 which in turn depict the areal distribution of surficial deposits in the study area.

The surficial deposit units were derived from the regional maps of Roed (1968, 1975), St-Onge (1967, 1970, 1972), and the author's interpretation of aerial photographs. Limited field checking allowed some refinement of the surficial deposit unit boundaries in the immediate vicinity of the Athabasca River valley. Much of the terminology for the various stratigraphic units discussed in Sections 3.2 and 3.3 is that established by Roed (1968, 1975) and St-Onge (1972). Some

TABLE 3-1

RELATIVE AGES	STRATIGRAPHIC UNITS	
Holocene Deposits	Organic Colluvial Eolian Alluvial Floodplain	
	Cordilleran	Laurentide
Pleistocene Deposits	Drystone Creek Till[1]	Deltaic Sands[2]
	Lower Hinton Valley Train Terrace[2]	Glaciolacustrine Sediments[2]
	Upper Hinton Valley Train Terrace[2] ⟷	Berland River Outwash Sands[2]
	Obed Till	↑
	Pedley Sediments[2]	
	Emerson Lakes Esker[2]	
	Marlboro Till ⟵ possibly time synchronous ⟶	Mayberne Till
		Marsh Creek Outwash[1]
	'Early' Cordilleran Till[1]	
Pre-Quaternary-Bedrock	Brazeau Formation	Paskapoo Formation

[1]Stratigraphic positioning of these units derived from Roed (1968, 1975)

[2]Units developed during the recessional phases of the previous glacial unit

Source: Roed (1968, 1975); St-Onge (1972); and author's field observations.

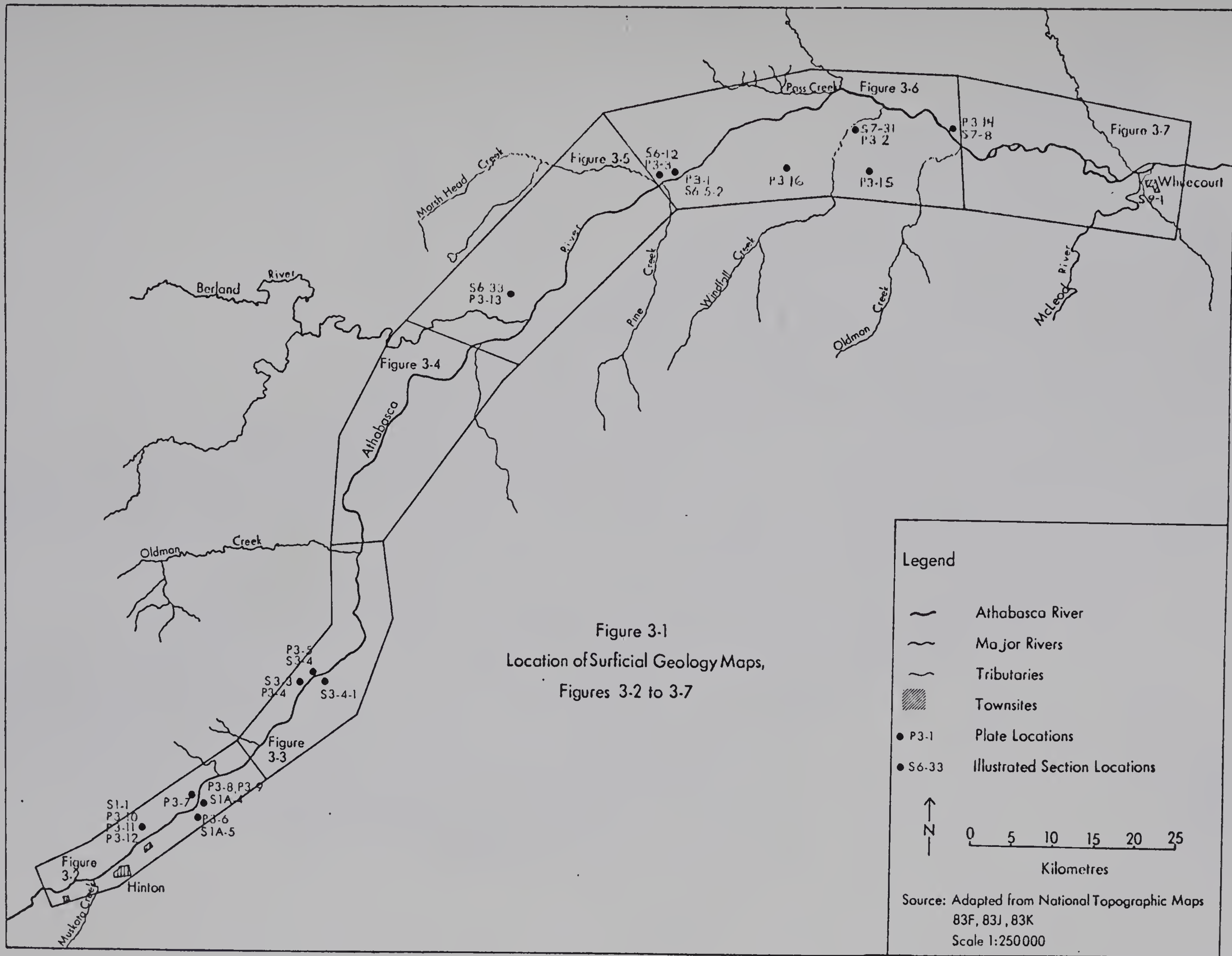


Figure 3-1
Location of Surficial Geology Maps,
Figures 3-2 to 3-7

LEGEND

UNIT	SUBUNIT	MORPHOLOGICAL DESCRIPTION	SURFICIAL MATERIAL
Glacial Deposits			
		Hummocky moraine	Till; mixture of cobbles, sand, silt and clay.
Glaciofluvial Deposits			
	1	Outwash fans or plains; some poorly defined channels	Sand and gravel; undifferentiated.
	2	Meltwater channel deposits	Terrace gravels and sand; some sand lenses.
Ice-contact Deposits			
	1	Esker complex	Sand and gravel; inclusions of till, silt and clay in some areas.
	2	Kame complex	Sand and gravel; inclusions of till, silt and clay.
Glaciolacustrine Deposits			
	1	Lacustrine plains	Clay, silt and fine sand; laminated
	2	Deltas, partly re-worked to form inactive eolian deposits	Medium to fine grained sands; dunes common.
Alluvial Deposits			
	1	Present channel flood-plain	Gravels, sand, silt and some clays.
	2	Alluvial terraces	Gravels, sand, silt and some clays.
Colluvial Deposits			
		Various mass movement forms	Mixed and weathered bed-rock, till, sands and silts
Organic Deposits			
		Peat bogs	Fine silt, clay and muck.

* 4-2	Illustrated section locations
R	Bedrock exposures
	Break in slope
	Glacial meltwater channel
	Course of abandoned stream channel
	Drumlin
	Groove or fluting
	Esker ridge
	Dunes; Parabolic, Linear
	Geomorphic boundary; Defined, Inferred
3500	Contour (feet); (1 foot = 0.304 meters) C.I. = 250 feet

Source: Roed (1968, 1975); St-Onge (1972); and author's field observations.

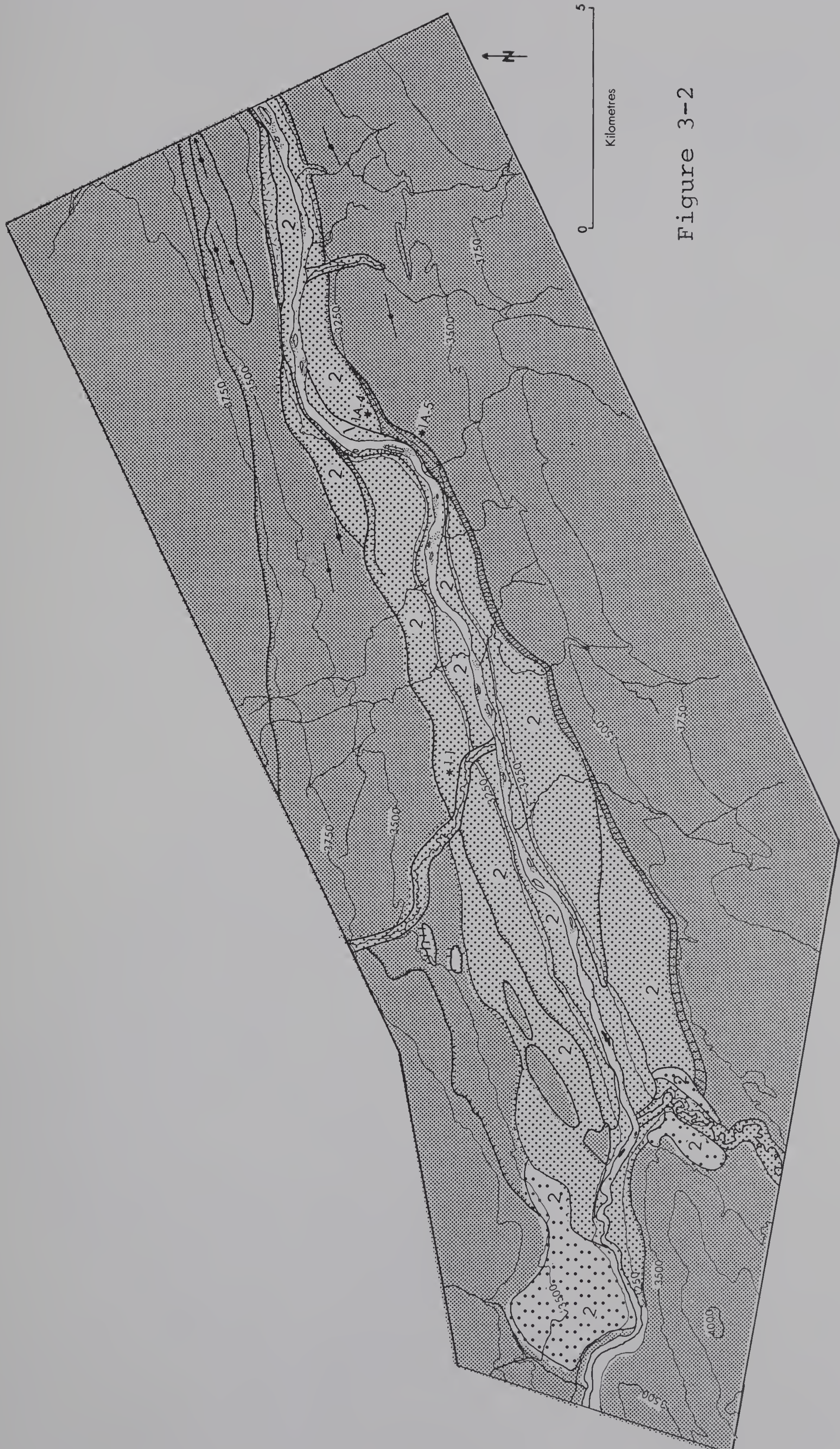


Figure 3-2

LEGEND

UNIT	SUBUNIT	MORPHOLOGICAL DESCRIPTION	SURFICIAL MATERIAL
Glacial Deposits			
		Hummocky moraine	Till; mixture of cobbles, sand, silt and clay.
Glaciofluvial Deposits			
	1	Outwash fans or plains; some poorly defined channels	Sand and gravel; undifferentiated.
	2	Meltwater channel deposits	Terrace gravels and sand; some sand lenses.
Ice-contact Deposits			
	1	Esker complex	Sand and gravel; inclusions of till, silt and clay in some areas.
	2	Kame complex	Sand and gravel; inclusions of till, silt and clay.
Glaciolacustrine Deposits			
	1	Lacustrine plains	Clay, silt and fine sand; laminated
	2	Deltas, partly re-worked to form inactive eolian deposits	Medium to fine grained sands; dunes common.
Alluvial Deposits			
	1	Present channel flood-plain	Gravels, sand, silt and some clays.
	2	Alluvial terraces	Gravels, sand, silt and some clays.
Colluvial Deposits			
		Various mass movement forms	Mixed and weathered bed-rock, till, sands and silts
Organic Deposits			
		Peat bogs	Fine silt, clay and muck.

4-2	Illustrated section locations
R	Bedrock exposures
	Break in slope
	Glacial meltwater channel
	Course of abandoned stream channel
	Drumlin
	Groove or fluting
	Esker ridge
	Dunes; Parabolic, Linear
	Geomorphic boundary; Defined, Inferred
3500	Contour (feet); (1 foot = 0.304 meters) C.I. = 250 feet

Source: Roed (1968, 1975); St-Onge (1972); and author's field observations.

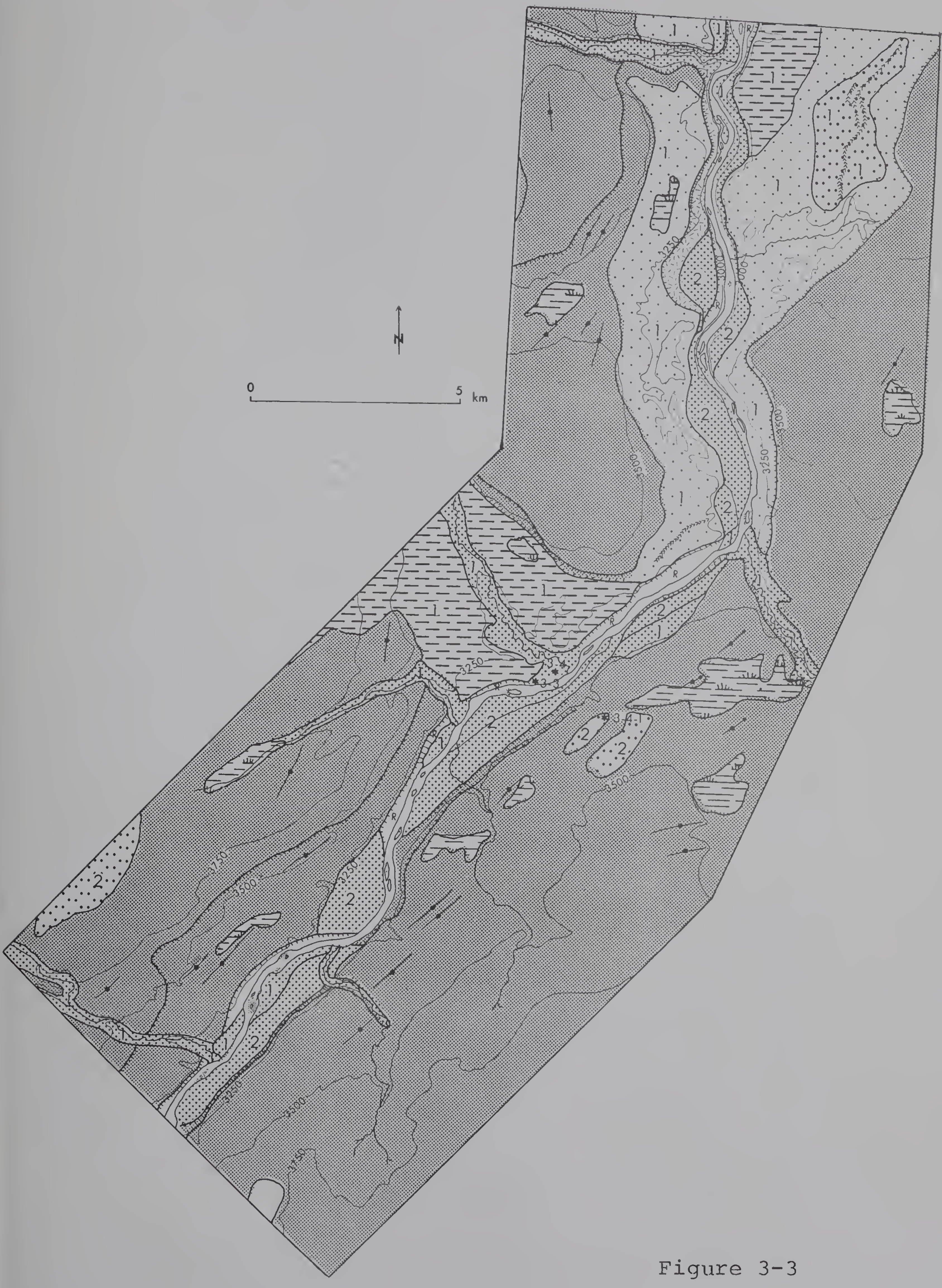


Figure 3-3

LEGEND

UNIT	SUBUNIT	MORPHOLOGICAL DESCRIPTION	SURFICIAL MATERIAL
Glacial Deposits			
		Hummocky moraine	Till; mixture of cobbles, sand, silt and clay.
Glaciofluvial Deposits			
	1	Outwash fans or plains; some poorly defined channels	Sand and gravel; undifferentiated.
	2	Meltwater channel deposits	Terrace gravels and sand; some sand lenses.
Ice-contact Deposits			
	1	Esker complex	Sand and gravel; inclusions of till, silt and clay in some areas.
	2	Kame complex	Sand and gravel; inclusions of till, silt and clay.
Glaciolacustrine Deposits			
	1	Lacustrine plains	Clay, silt and fine sand; laminated
	2	Deltas, partly re-worked to form inactive eolian deposits	Medium to fine grained sands; dunes common.
Alluvial Deposits			
	1	Present channel flood-plain	Gravels, sand, silt and some clays.
	2	Alluvial terraces	Gravels, sand, silt and some clays.
Colluvial Deposits			
		Various mass movement forms	Mixed and weathered bed-rock, till, sands and silts
Organic Deposits			
		Peat bogs	Fine silt, clay and muck.

* 4-2	Illustrated section locations
R	Bedrock exposures
	Break in slope
	Glacial meltwater channel
	Course of abandoned stream channel
	Drumlin
	Groove or fluting
	Esker ridge
	Dunes; Parabolic, Linear
	Geomorphic boundary; Defined, Inferred
3500	Contour (feet); (1 foot = 0.304 meters) C.I. = 250 feet

Source: Roed (1968, 1975); St-Onge (1972); and author's field observations.

Figure 3-4

LEGEND

UNIT	SUBUNIT	MORPHOLOGICAL DESCRIPTION	SURFICIAL MATERIAL
Glacial Deposits			
		Hummocky moraine	Till; mixture of cobbles, sand, silt and clay.
Glaciofluvial Deposits			
	1	Outwash fans or plains; some poorly defined channels	Sand and gravel; undifferentiated.
	2	Meltwater channel deposits	Terrace gravels and sand; some sand lenses.
Ice-contact Deposits			
	1	Esker complex	Sand and gravel; inclusions of till, silt and clay in some areas.
	2	Kame complex	Sand and gravel; inclusions of till, silt and clay.
Glaciolacustrine Deposits			
	1	Lacustrine plains	Clay, silt and fine sand; laminated
	2	Deltas, partly re-worked to form inactive eolian deposits	Medium to fine grained sands; dunes common.
Alluvial Deposits			
	1	Present channel flood-plain	Gravels, sand, silt and some clays.
	2	Alluvial terraces	Gravels, sand, silt and some clays.
Colluvial Deposits			
		Various mass movement forms	Mixed and weathered bed-rock, till, sands and silts
Organic Deposits			
		Peat bogs	Fine silt, clay and muck.

* 4-2	Illustrated section locations
R	Bedrock exposures
	Break in slope
	Glacial meltwater channel
	Course of abandoned stream channel
	Drumlin
	Groove or fluting
	Esker ridge
	Dunes; Parabolic, Linear
	Geomorphic boundary; Defined, Inferred
3500	Contour (feet); (1 foot = 0.304 meters) C.I. = 250 feet

Source: Roed (1968, 1975); St-Onge (1972); and author's field observations.

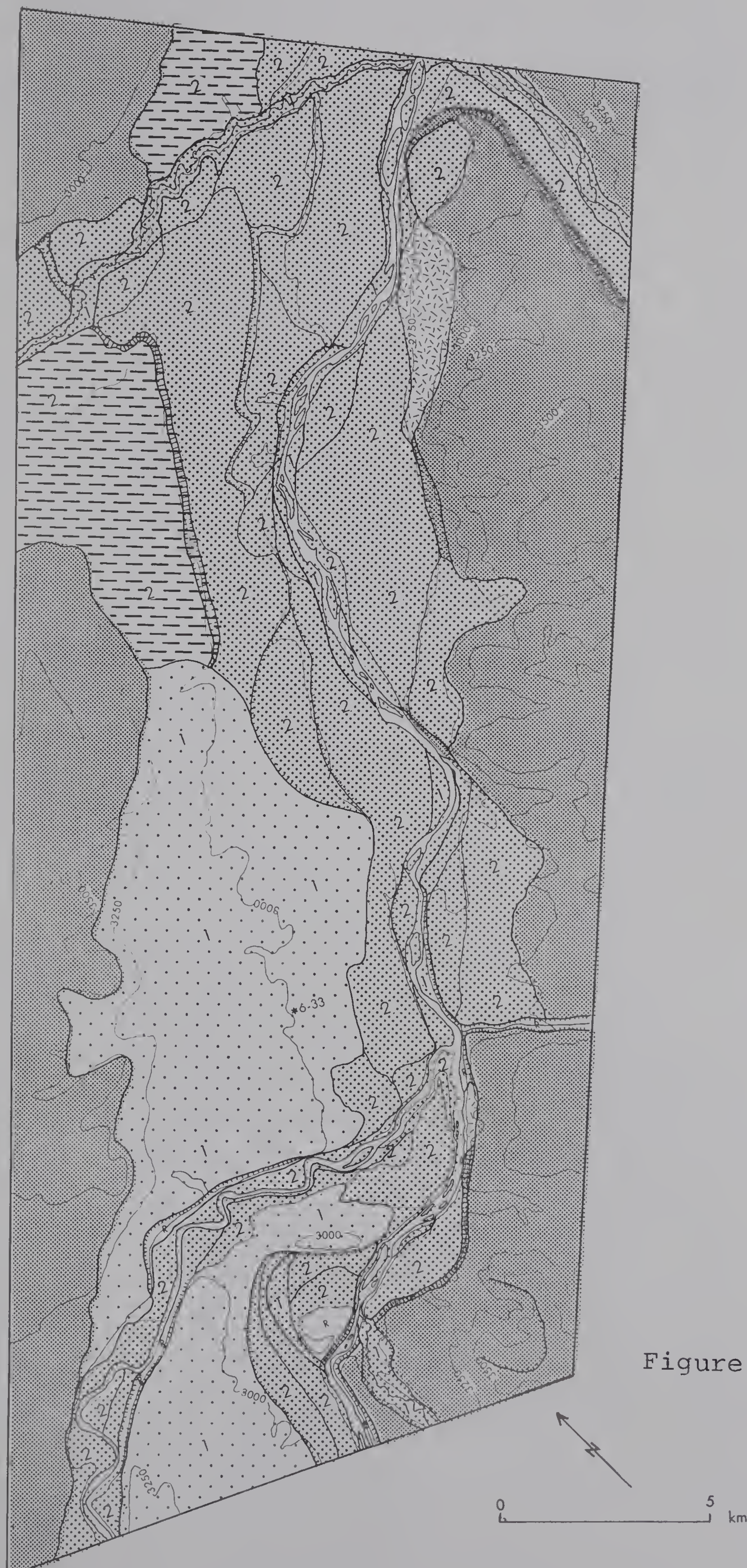


Figure 3-5

LEGEND

UNIT	SUBUNIT	MORPHOLOGICAL DESCRIPTION	SURFICIAL MATERIAL
Glacial Deposits			
		Hummocky moraine	Till; mixture of cobbles, sand, silt and clay.
Glaciofluvial Deposits			
	1	Outwash fans or plains; some poorly defined channels	Sand and gravel; undifferentiated.
	2	Meltwater channel deposits	Terrace gravels and sand; some sand lenses.
Ice-contact Deposits			
	1	Esker complex	Sand and gravel; inclusions of till, silt and clay in some areas.
	2	Kame complex	Sand and gravel; inclusions of till, silt and clay.
Glaciolacustrine Deposits			
	1	Lacustrine plains	Clay, silt and fine sand; laminated
	2	Deltas, partly re-worked to form inactive eolian deposits	Medium to fine grained sands; dunes common.
Alluvial Deposits			
	1	Present channel flood-plain	Gravels, sand, silt and some clays.
	2	Alluvial terraces	Gravels, sand, silt and some clays.
Colluvial Deposits			
		Various mass movement forms	Mixed and weathered bed-rock, till, sands and silts
Organic Deposits			
		Peat bogs	Fine silt, clay and muck.

* 4-2	Illustrated section locations
R	Bedrock exposures
	Break in slope
	Glacial meltwater channel
	Course of abandoned stream channel
	Drumlin
	Groove or fluting
	Esker ridge
	Dunes; Parabolic, Linear
	Geomorphic boundary; Defined, Inferred
3500	Contour (feet); (1 foot = 0.304 meters) C.I. = 250 feet

Source: Roed (1968, 1975); St-Onge (1972); and author's field observations.

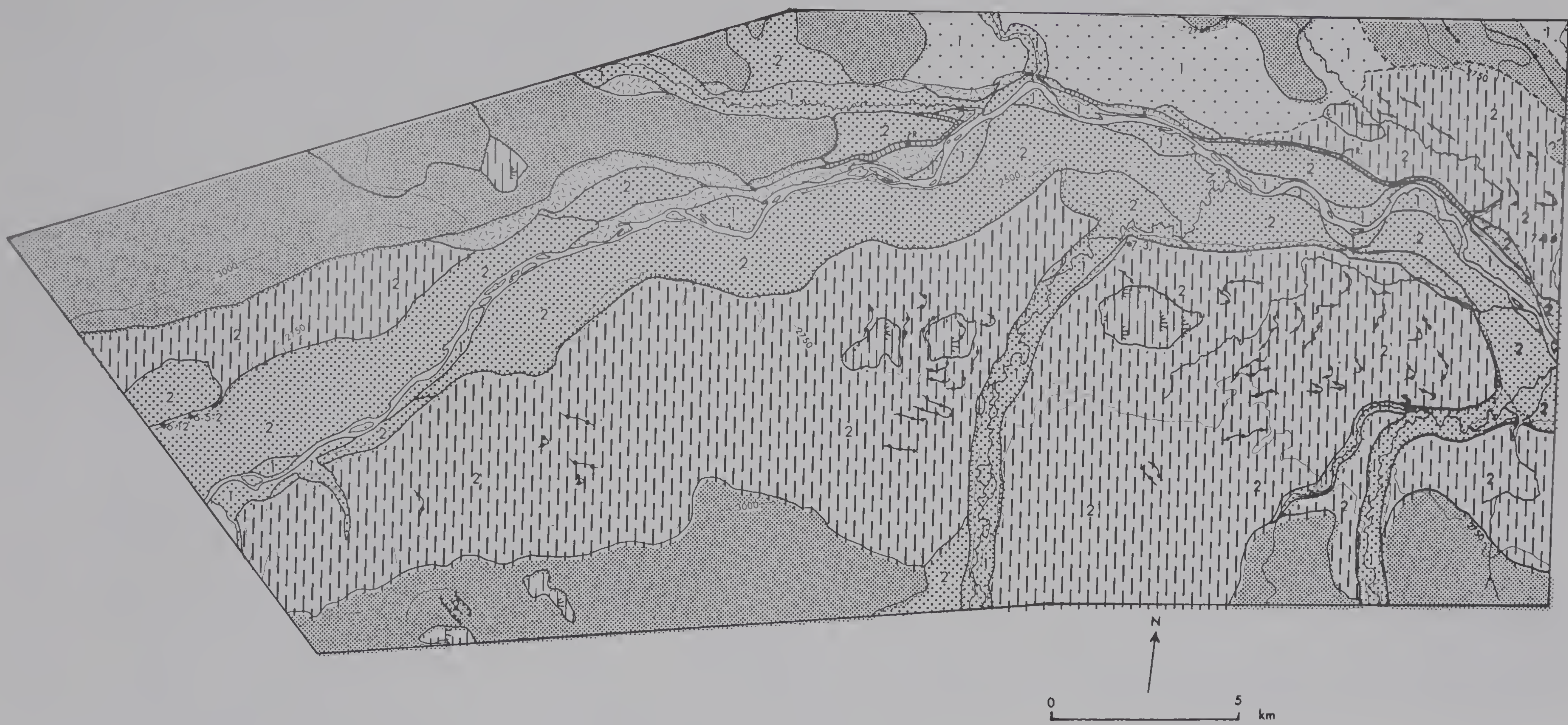


Figure 3-6

LEGEND

UNIT	SUBUNIT	MORPHOLOGICAL DESCRIPTION	SURFICIAL MATERIAL
Glacial Deposits			
		Hummocky moraine	Till; mixture of cobbles, sand, silt and clay.
Glaciofluvial Deposits			
	1	Outwash fans or plains; some poorly defined channels	Sand and gravel; undifferentiated.
	2	Meltwater channel deposits	Terrace gravels and sand; some sand lenses.
Ice-contact Deposits			
	1	Esker complex	Sand and gravel; inclusions of till, silt and clay in some areas.
	2	Kame complex	Sand and gravel; inclusions of till, silt and clay.
Glaciolacustrine Deposits			
	1	Lacustrine plains	Clay, silt and fine sand; laminated
	2	Deltas, partly re-worked to form inactive eolian deposits	Medium to fine grained sands; dunes common.
Alluvial Deposits			
	1	Present channel flood-plain	Gravels, sand, silt and some clays.
	2	Alluvial terraces	Gravels, sand, silt and some clays.
Colluvial Deposits			
		Various mass movement forms	Mixed and weathered bed-rock, till, sands and silts
Organic Deposits			
		Peat bogs	Fine silt, clay and muck.

* 4-2	Illustrated section locations
R	Bedrock exposures
	Break in slope
	Glacial meltwater channel
	Course of abandoned stream channel
	Drumlin
	Groove or fluting
	Esker ridge
	Dunes; Parabolic, Linear
	Geomorphic boundary; Defined, Inferred
3500	Contour (feet); (1 foot = 0.304 meters) C.I. = 250 feet

Source: Roed (1968, 1975); St-Onge (1972); and author's field observations.

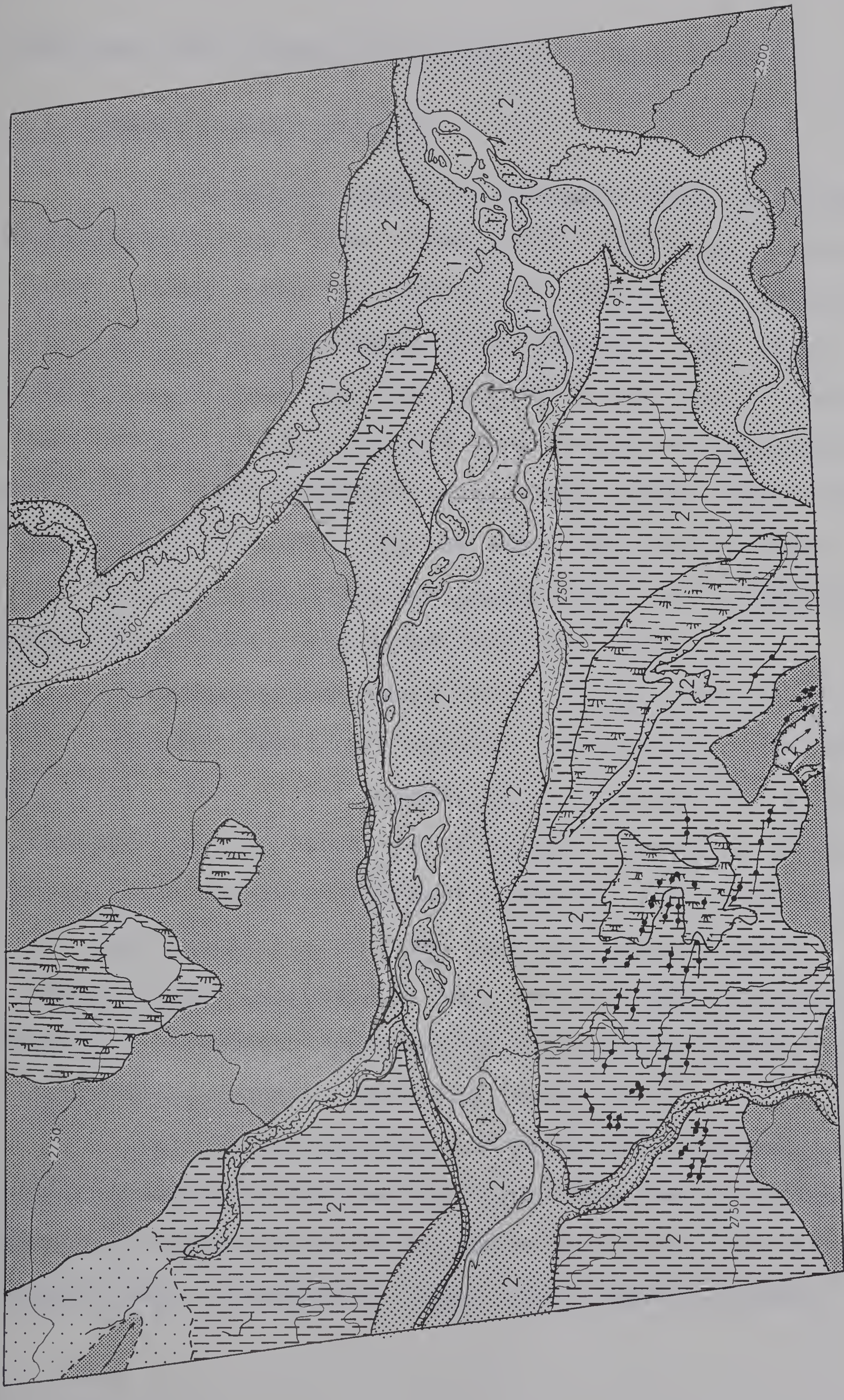


Figure 3-7

additional unit terms are introduced in this study.

3.2 General Glacial History

A complicated history of glacial events is evident for the Athabasca River area. Roed (1968, 1975), identified an 'Early' Cordilleran ice advance and termed the first Laurentide ice intrusion the Marsh Creek advance. The 'Early' Cordilleran advance may have reached as far east as Nosehill Creek before receding into the area west of a line connecting the Wildhay River and Marsh Creek (Roed, 1975). The Marsh Creek glacier advanced into the area from the northeast, at least as far as the above-mentioned line. As it retreated the Marsh Creek outwash was deposited (Roed, 1975).

The next major glacial event was the advance of both the Cordilleran (Marlboro) glacier and the Laurentide (Edson) ice sheet. Roed (1975) suggested that the Marlboro glacier, spreading over the Athabasca tablelands, came into direct contact with the Edson ice mass along the line shown in Figure 3-8. The northern part of the Marlboro glacier became stagnant during coalescence, the southern part continuing to flow in a southeasterly direction between the Laurentide ice and the Rocky Mountain foothills (Roed, 1975).

Glaciolacustrine deposits began to accumulate in the lower Oldman Creek area as the Cordilleran and Laurentide ice sheets initiated recession to the southwest and northeast respectively. The Emerson Lakes esker complex probably

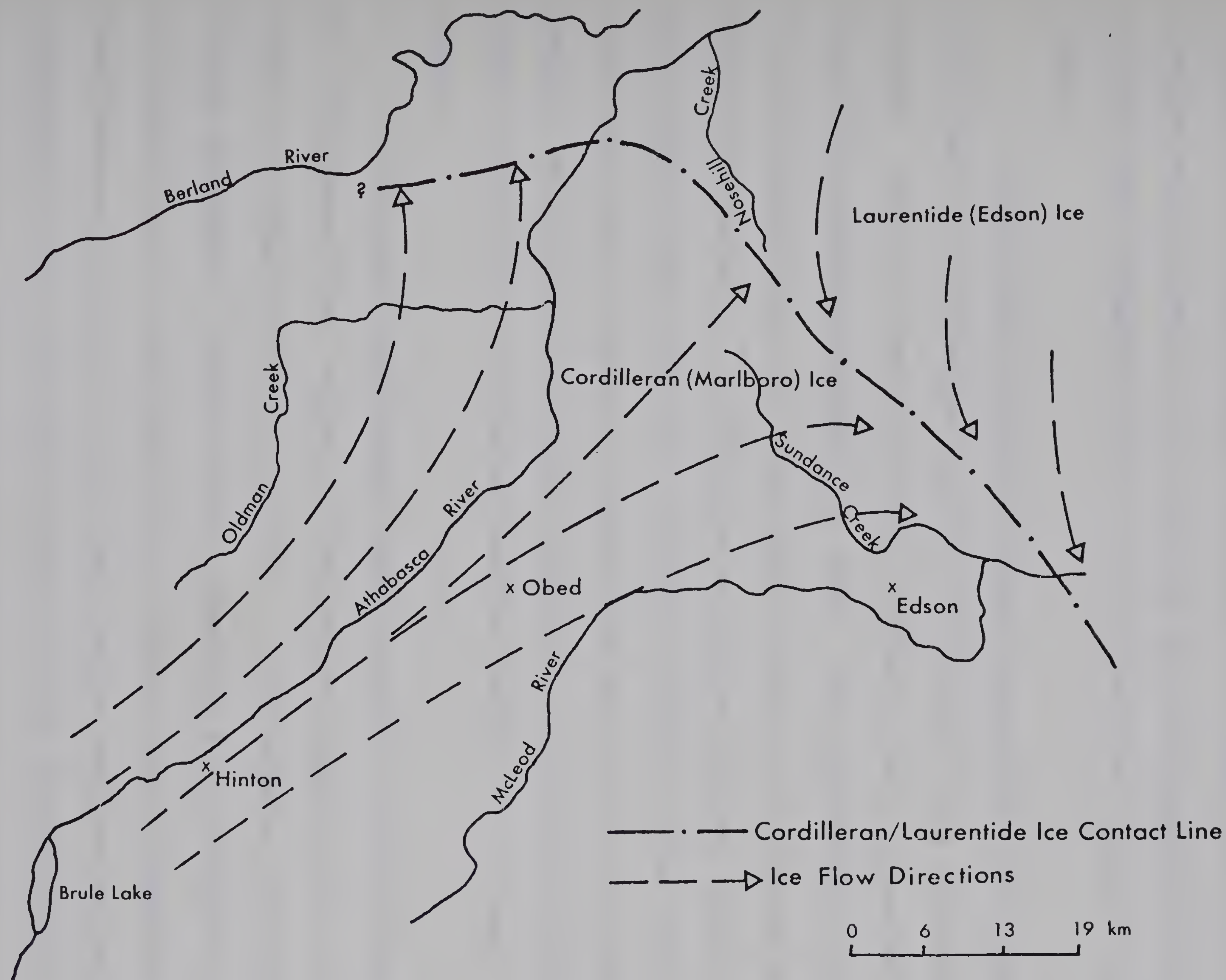


Figure 3-8. Suggested line of contact between the Cordilleran and Laurentide ice sheets, (after Roed, 1975).

developed while the two ice masses were still in close proximity (Figure 3-3). Cordilleran meltwater in transit from the esker complex was blocked to the north by the Laurentide ice and spilled down Sundance Creek as an ice marginal channel (Roed, 1968).

The Marlboro glacier receded westward to the Brûle Lake area where it is thought to have stabilized for a considerable period. At this time the Pedley sediments were deposited within the Athabasca benchland zone. The interbedded and discontinuous clay, silt, sand and gravel lenses making up the Pedley Sediments are indicative of braided channel deposition. This depositional sequence, in turn, attests to the channel conditions which existed in the valley during this aggradational period, as sediment/discharge conditions varied with fluctuations in meltwater discharge from the nearby Cordilleran ice front.

The later Cordilleran (Obed) glacier advanced down the Athabasca River valley to the present mapped limits of Obed till (Roed, 1975; Figure 3-3). This advance overrode the Pedley sediments. Kame deposits (Section 3-4-1, Figure 3-3), recognized as ice distintegration features, were deposited near the eastern limit of the Obed glacier and are probably indicative of the very early stages of Obed glacier retreat from this area. Eventually, the Obed glacier receded and stabilized in the area of Muskata Creek (Figure 3-1). During this still-stand phase the upper Hinton valley train terrace

sediments were deposited. With the subsequent retreat of the Obed glacier excess meltwaters from the glacier entrenched the upper valley train deposits and underlying materials until such time that the Obed glacier experienced a second readvance and/or still-stand phase. While the exact ice frontal position of this phase is uncertain it was probably during this period that the lower Hinton valley train sediments were deposited. Further wasting of the Obed glacier again increased meltwater flows down the Athabasca River. The increased flow dissected the previously deposited sediments to form the lower valley train terrace (Stene, 1966).

Whether the major retreat phases of the Cordilleran and Laurentide ice masses were time-synchronous or not is unknown. However, after the deposition of an outwash complex in the lower Berland River area Laurentide ice continued to waste towards the northeast. Meltwater directed along the Laurentide ice front, then lying south of Fox Creek, carved two channels across a nearby bedrock ridge north of the Athabasca River valley. The channel to the southeast of Fox Creek is presently occupied by Pass Creek and the second channel, 30 kilometers to the west, is occupied by a segment of the Little Smoky River (St-Onge, 1972). Meltwater was dammed to the east by the Laurentide ice sheet and accumulated in the Athabasca River valley in the vicinity of Windfall Creek. This meltwater formed Glacial Lake Windfall (St-Onge, 1972). A large delta complex was constructed by both the Athabasca River and the

Little Smoky River spillway at the western end of Glacial Lake Windfall. The elevation of this lake lay between 870 and 840 meters (Figure 3-9). It drained to the south into Glacial Lake Peers through a spillway 16 kilometers southwest of Whitecourt.

Continued retreat of the Laurentide ice sheet resulted in the recession of Glacial Lake Windfall. Meltwater from the northwest continued to flow through Pass Creek into the Athabasca River valley and Glacial Lake Wildwood (Figure 3-10). Lake Wildwood extended from the large delta complex near Windfall, at an elevation of approximately 830 meters, southeast of Thorsby. Further retreat of the Laurentide ice to a position approximately 10 kilometers west of Blue Ridge allowed the Athabasca River and Pass Creek drainage systems to build a large sandy delta into the northwestern portion of Glacial Lake Leduc, near Whitecourt, at an elevation of 750 to 720 meters (Figure 3-11).

Several C^{14} dates were obtained by St-Onge (1972) from organic materials contained in sediments of his study area. These indicate that locally the Laurentide deglaciation may have been relatively rapid and the dates give minimum limiting estimates of times of deglaciation (Figure 3-12). The lowering and eventual disappearence of proglacial lakes along the river exposed large tracts of deltaic sands and finer lacustrine sediments. Before vegetation could effectively colonize these

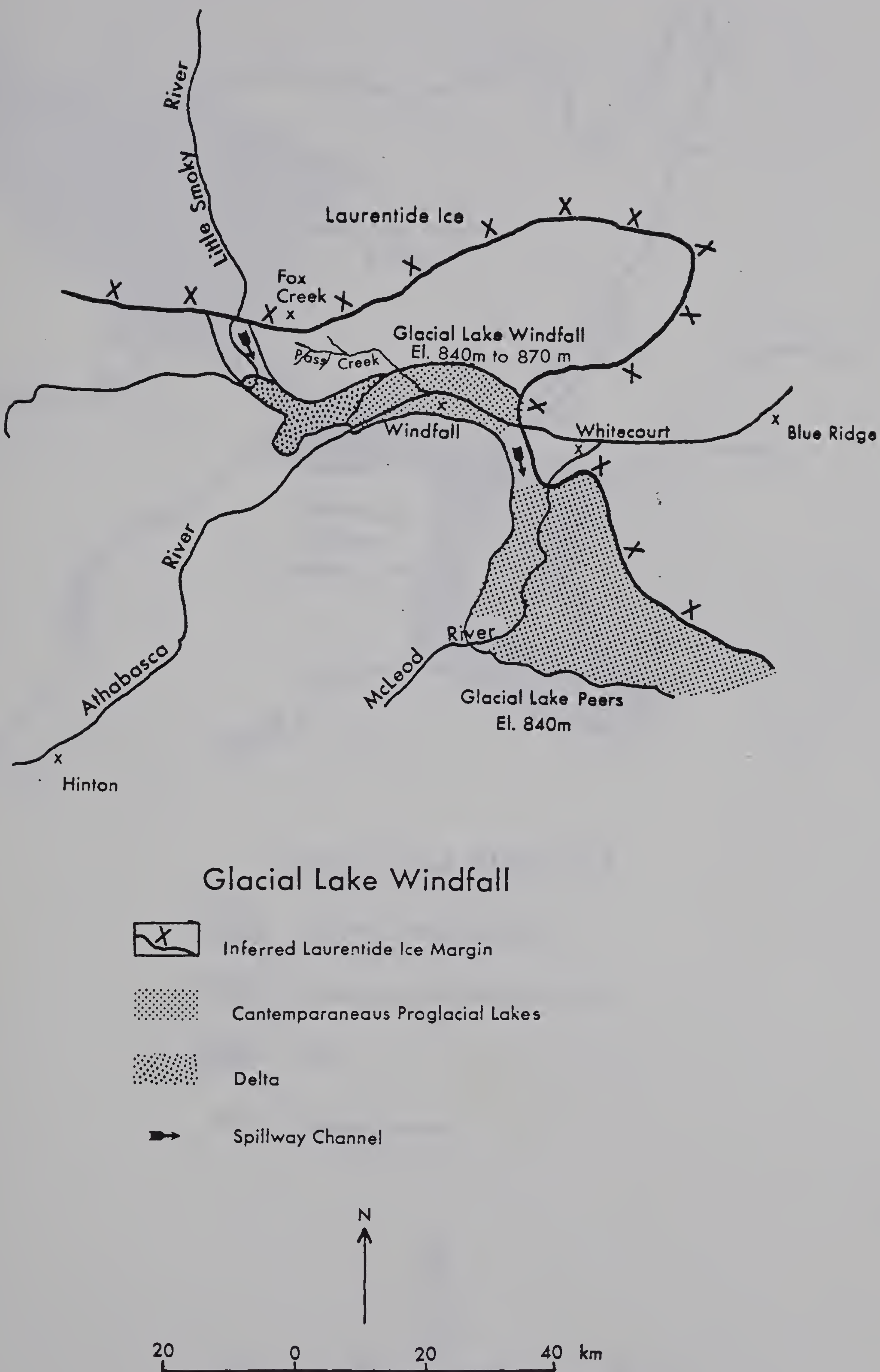


Figure 3-9. Location of Glacial Lake Windfall, (after St-Onge, 1972).

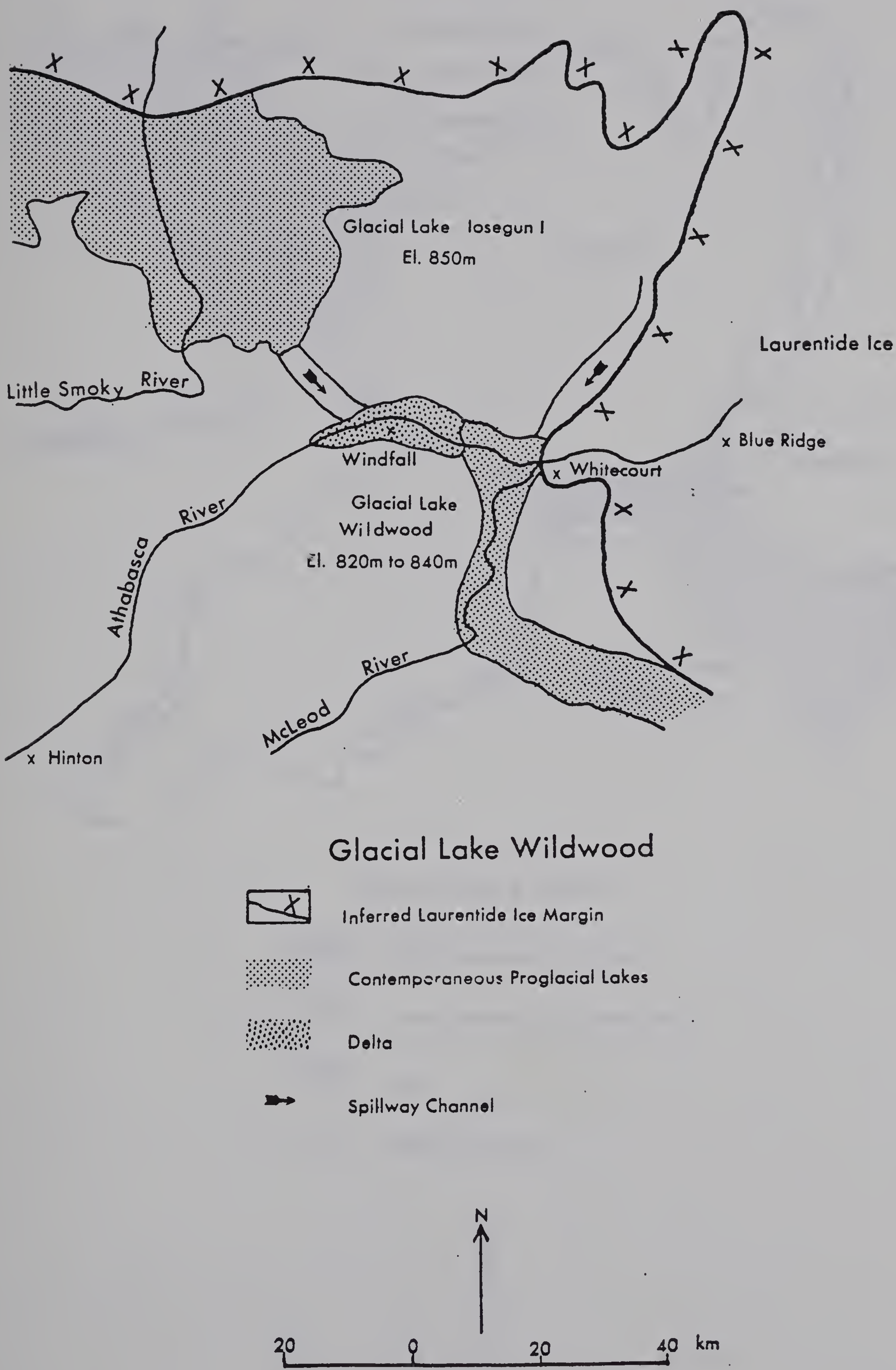


Figure 3-10. Location of Glacial Lake Wildwood, (after St-Onge, 1972).

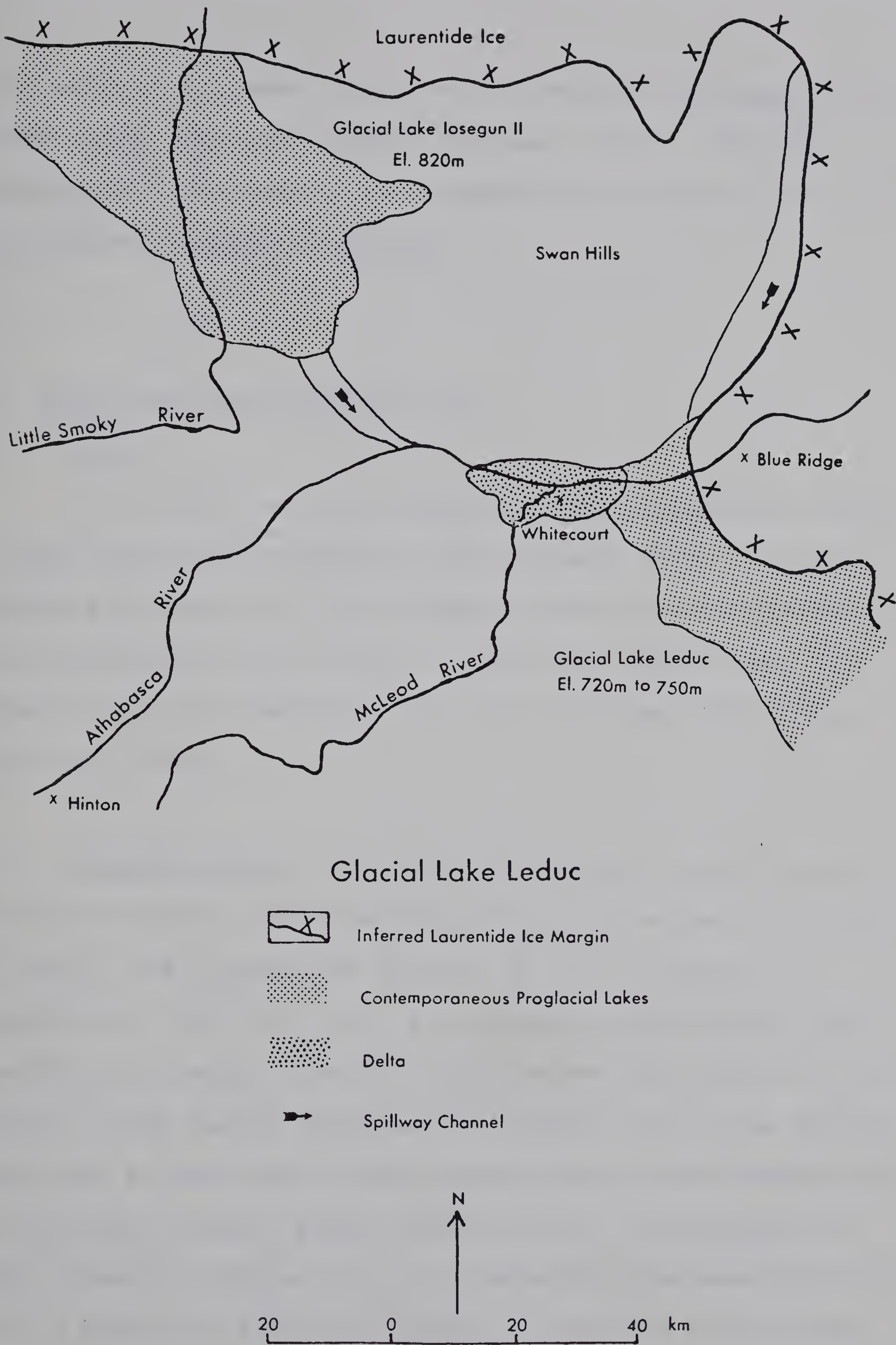


Figure 3-11. Location of Glacial Lake Leduc, (after St-Onge, 1972).

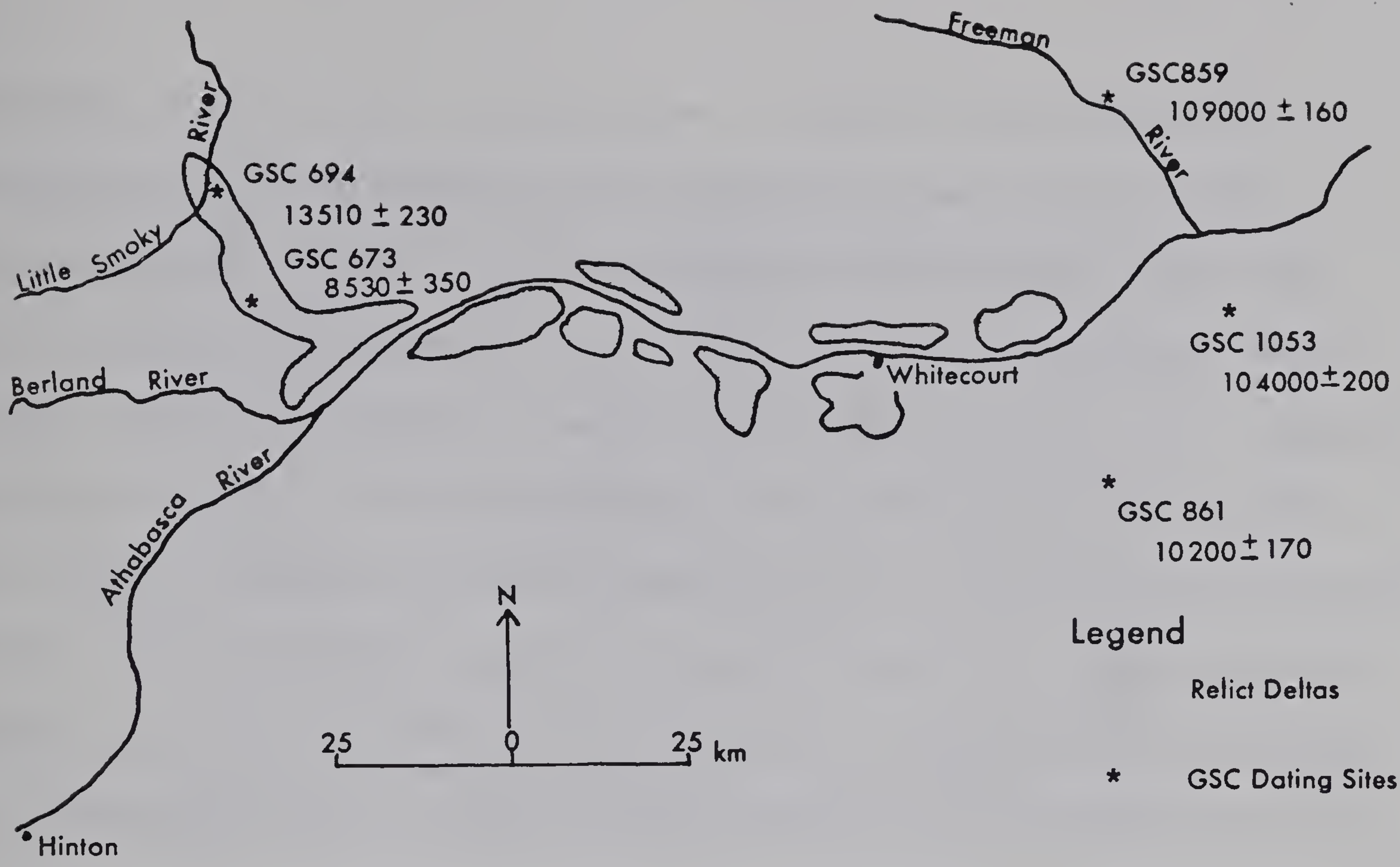
newly exposed surfaces strong winds reworked the deposits and formed large sand dune fields (St-Onge, 1972). The disappearence of the lakes also allowed the Athabasca River to progressively erode its valley.

3.3 Major Surficial Deposit Units

3.3.1 Tills

Tills are the most widespread glacial deposits found in the upland areas bordering the Athabasca River. In the immediate proximity of the Athabasca River three till units were identified by the author. These were termed the; (1) Mayberne till; (2) Marlboro till; and (3) Obed till; after Roed (1968, 1975).

Mayberne Till. This till represents the principal surficial deposit of the borderlands in the eastern sector of the area. The illustrated section is 6-5-2 (Figure 3-6, Appendix A). The till unit is 2.5 meters thick at the type locality, but ranges upward to thicknesses of 6 meters in the Windfall Creek valley (Section 7-3, Figure 3-6). The till is light grey to dark grey brown, plastic when moist, moderately to very stony, with a pebble mode of 2 to 3 centimeters in clast diameters (Plate 3-1). It contains occasional boulders up to 2 meters in diameter (Plate 3-2) but has a clay loam



GSC 673	8,530±350	Fresh water gastropod shells in cross-bedded sands; S. Kaybob Oilfield
GSC 861	10,200±170	gastropods from thin lenses of silty sand in thick deposits of lacustrine silts at a depth of 100 centimeters; 10.5 kilometers east of Greencourt
GSC 1053	10,400±200	organic material overlain by silty clay at a depth of 476 centimeters; Clear Lake, elevation 700 meters
GSC 859	10,900±160	wood fragment below till at a depth of 450 centimeters; Freeman River
GSC 694	13,510±230	large shells; Little Smoky River

Figure 3-12. C^{14} dates from organic materials in the proximity of the Athabasca River, (after St-Onge, 1972).

matrix, and is well consolidated. Clasts are rounded to subangular, predominantly metaquartzites and crystalline metamorphics, with a low percentage of sandstone, siltstone and ironstone fragments. This till overlies the shale/sandstone Paskapoo Formation and, in some localities, early buried valley gravels. Buried valleys were defined by Roed (1968), as valleys which have been formed mainly by stream erosion and later infilled with glacial or other materials. The gravels which occurred near or at the bottom of these buried valleys, he referred to as Buried Valley Gravels. No attempt has been made to distinguish the age of these gravels. They do not necessarily have to predate glacial events since the occurrence of interglacial buried valleys would also be possible. As well, these deposits may be representative of glacial outwash laid down prior to ice advance. Although no exposure showing its contact with the Paskapoo Formation was found by the author one exposure of the tills contact with the buried valley gravels was located (Plate 3-3). This section occurs on the south facing bank of the Athabasca River (Section 6-12, Figure 3-6) and is exposed along an abandoned road cut. Here the buried valley gravels form a bed 3 to 4 meters thick. The gravels, in a coarse sandy matrix, have no apparent bedding structure, and modal diameters of approximately 5 to 10 centimeters. The clasts are mainly rounded to subrounded, quartzites,

Plate 3-1. Illustrated Mayberne till section 6-5-2.

Plate 3-2. Illustrated Mayberne till section 7-31.

Plate 3-3. Mayberne till overlying buried valley gravels, section 6-12.

limestones and sandstones.

Marlboro Till. Marlboro till is exposed at only a few localities in the study area. At the illustrated section, 3-4, (Figure 3-3, Appendix A) 3 meters of Marlboro till overly sandstone of the Paskapoo Formation (Plate 3-4). The till is pale olive to light brown, very slightly plastic when moist, moderately stony, with a pebble mode of approximately 2.5 centimeters in diameter. The till matrix varies from silty, sandy clay to clay loam, and is not highly consolidated. Pebbles are subrounded to subangular quartzites, limestones, ortho-quartzites, meta-quartzites and sandstones (Plate 3-5). Contacts between the Marlboro till and more recent, overlying deposits were not found by the author. Roed (1968, 1975), however, determined that in the Athabasca Benchlands and river valley areas Marlboro till lies directly beneath Cordilleran outwash deposits (Pedley Sediments) which in turn are overlain by Obed till. Elsewhere in the Pedley/Obed area he found that Obed till rests directly on Marlboro till.

Obed Till. The Obed till is the most extensive surficial deposit in the western part of the study area. The illustrated section, 1A-5, (Figure 3-2, Appendix A), is located on the north facing bank of the Athabasca River. The till bed at this locality is approximately 3 meters thick, the till thickness varying at other places from 1.5 to 5 meters. The till is olive brown to light brown, slightly plastic, moderately

Plate 3-4. Marlboro till overlying bedrock, section 3-4.

Plate 3-5. Illustrated Marlboro till section 3-3.

to very stony, with a pebble diameter mode of approximately 3 centimeters (Plate 3-6). Boulders up to 45 centimeters in diameter are also included (Plate 3-7). The till has a high carbonate content with clasts composed primarily of limestones, sandstones, quartzites and some conglomerates. The till has a sandy silt/clay matrix, and is moderately consolidated.

3.3.2 Other Stratigraphic Units

Other stratigraphic units recognized in the area include Pleistocene subsurface outwash, glaciofluvial, glacio-lacustrine and non-glacial Holocene deposits.

Subsurface Outwash Deposits. Subsurface outwash deposits are recognized by their stratigraphic position between tills and by their structure, which together suggest these sediments were deposited in a moving water environment, during a period of glacial still-stand or recession. These subsurface outwash deposits may be more correctly interpreted as interstadial or interglacial deposits.

The only subsurface outwash deposits so far identified in the study area are the Pedley Sediments. These sediments do not form a surface deposit but occur stratigraphically between two tills of different ages. The deposits are widespread in the Athabasca borderlands, extending from Hinton as far east as Obed (Roed, 1968). The illustrated section (Figure 3-2, Appendix A) is exposed along a north

Plate 3-6. Illustrated Obed till section 1A-5.

Plate 3-7. Illustrated Obed till section 1A-1.

facing slope of the Athabasca River. Here the unit is approximately 15 meters thick, and is overlain by Obed till (Plate 3-8).

The character of the Pedley Sediments is highly variable. The modal clast diameters in the unit vary from 20 to 50 centimeters. Fine gravel lenses are interspersed with coarser gravel beds, suggestive of braided river deposition. Pebbles are rounded to subrounded, dominated by locally derived sandstones, limestones and quartzites with minor traces of chert.

Throughout the illustrated section 1A-4 a series of stratified sand lenses also occur (Plate 3-9A and 3-9B). The development of these sand lenses may be attributed to the rapidly varying channel conditions which existed during the deposition of this outwash sequence. In a mixed gravel-sand bed, braided channel transverse and linguoid sand bars occur in the relatively broad, shallow channels at low flow velocities, varying in plan view from obliquely rectilinear to convex downstream (Allen, 1970). Following the development of these bar forms throughout the channel network, a change in channel hydraulics (resulting from a change in sediment load, discharge or both) probably destroyed many of these bar forms by erosion, while others were buried and preserved beneath a subsequent gravel fill.

Plate 3-8. Pedley sediments overlain by Obed till.

Plate 3-9A. Illustrated Pedley
sediments section lA-4.

Plate 3-9B. Sand lenses within the Pedley sediments
section lA-4.

Glaciofluvial Deposits. Glaciofluvial deposits in the area include those of the Hinton Valley Train Terraces and the Berland River outwash sands complex. The Hinton terraces are valley train gravel deposits in the Hinton/Obed district. The illustrated section, 1-1, (Figure 3-2, Appendix A) is located in a gravel pit on the north side of the Athabasca River at Hinton. The terrace gravels at this locality are approximately 8 meters thick, representing deposits of the lower valley train. As for the Pedley Sediments, the character of the Hinton terrace materials is highly variable. Clasts are rounded to well rounded, primarily limestones, quartzites, with some sandstones and conglomerates in a medium to coarse sand matrix. The unit is poorly consolidated and modal clast diameters range from a few millimeters to 50 centimeters. The sand and gravel deposits display sections of contorted, cross-bedded, sand lenses and a series of fining upward cycles in the gravels (Plate 3-10). Gravels in the lower unit are more poorly sorted (Plate 3-11) than those of the upper unit (Plate 3-12). A further discussion of the Hinton terraces is contained in Chapter IV.

The Berland River outwash sands illustrated section, 6-33, (Figure 3-5, Appendix A) is located at a road cut on the north side of the Athabasca River. The sands of this locality are of variable thickness but average approximately 3 meters in depth. The unit consists of well consolidated fine sands and silts, with minor amounts of clay, overlain by coarse sands. The sand units are cross-stratified and parallel bedded

Plate 3-10. Illustrated section 1-1 of T_{2H} alluvium.

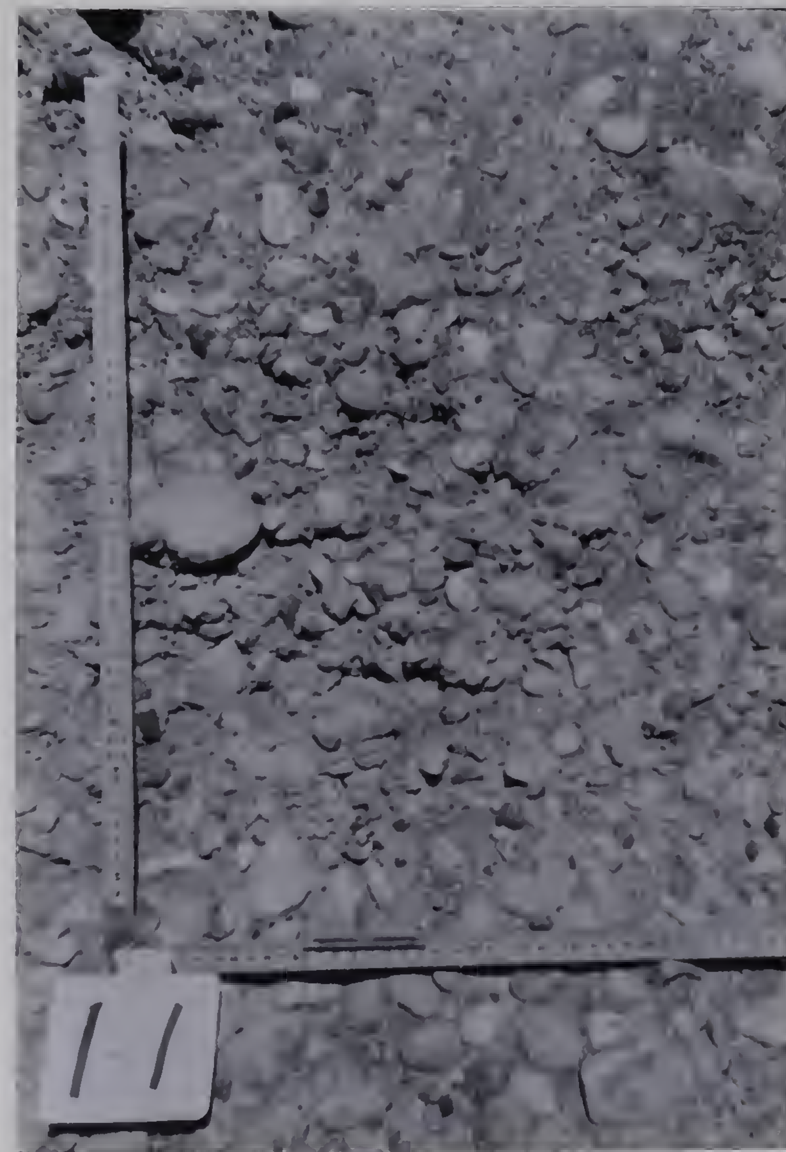


Plate 3-11. Illustrated section 1-1, lower alluvial unit.

Plate 3-12. Illustrated section 1-1, upper alluvial unit.

Plate 3-13. Illustrated Berland outwash section, 6-33.

(Plate 3-13) with some minor gravel lenses.

Glaciolacustrine Deposits. Glaciolacustrine deposits include lacustrine clays, silts and sands, and associated deltaic deposits. The sediments were deposited in proglacial lake environments which formed in front of the Laurentide ice sheet as it receded to the northeast. No exposures of proglacial lacustrine silt and clay deposits were identified in the study area. However several good exposures of proglacial deltaic deposits were identified.

Deltaic deposits are exposed at many localities in the borderland areas, east of the Berland River. At illustrated section 6-6 (Figure 3-6, Appendix A) approximately 2 meters of deltaic sands are exposed along a road cut, north of the Athabasca River. The sediments consist of medium to fine sands, with minor coal fragments. The unit displays large-scale, high angle, planar cross-strata, with thin, interbedded lenses of coal fragments (Plate 3-14).

Holocene Deposits. Deposits of non-glacial origin, and mainly of Holocene ages, include eolian, alluvial, colluvial and organic materials. Eolian dune fields predominate in close proximity to the deltaic deposits described earlier. The dune sediments were initially derived from the deltaic deposits formed along the Athabasca River. The dunes consist mainly of fine to very fine sands, which have formed parabolic dunes. The dune fields are now relatively stable because of binding by vegetation (Plate 3-15).

Plate 3-14. Illustrated deltaic
sand section 7-8.

Plate 3-15. Relict dune field.

Alluvial deposits are found along the Athabasca River, and its tributary streams, as low terrace and flood-plain deposits. These sediments are generally composed of gravels with a coarse sand matrix. Fine grained sands commonly overly gravel deposits in the low alluvial terraces. Further discussion of these deposits in contained in Chapter IV.

Colluvial deposits occur at various locations along the Athabasca River valley and its tributary streams, where steep valley walls predominate (Figure 3-7). Deposits include mixed, weathered bedrock and till, slump, soil creep and rock fall deposits and are thus composed of poorly sorted gravels, sands and silts.

Organic deposits occur throughout many low lying, poorly drained tableland areas bordering the Athabasca River (Figure 3-6). Deposits consist mainly of stone-free muck and peat, laid down in still-water bogs and marshes (Plate 3-16). These sediments typically overlie Laurentide tills and lacustrine sediments. The high clay content and compacted nature of these basal materials restrict the percolation of surface waters.

3.4 General Conclusions

The identification and stratigraphic assignment of the various surficial units, in particular the Pleistocene deposits, found along the Athabasca River, are critical when attempting to establish a connecting link between the Cordilleran and

Plate 3-16. Low lying marsh area responsible for the deposition of local organic deposits.

Laurentide glaciations of this area. Recognition and identification of the stratigraphic position of the various till units in the area provides the basis for developing a more comprehensive understanding of the number and extent of glacial advances into the region. More importantly, the recognition of glaciofluvial and glaciolacustrine deposits serves a duel purpose. First, the identification and stratagraphic position of these deposits provides the basis for tracing the retreat stages of the two glacier masses and understanding the sequence of changes each glacial system underwent. Related closely to this, and central to the main theme of this study, is the necessity to correlate remnant proglacial lake deposits with associated alluvial terraces. The distribution of these features within the valley reflects most closely the critical links between the two glacier systems.

Thus the proper recognition of each surficial deposit, its relative stratigraphic position, as well as its boundaries within the valley, provide the initial base for interpretation of the Pleistocene history of the area. As well, throughout much of the valley, Pleistocene and bedrock units were the principal sources of materials utilized for alluvial terrace deposition. Analysis of the clast composition of these various units is thus important. Central to this aspect is the occurrence of Canadian Shield clasts in terrace gravels, which would have been derived from the erosion of Laurentide till units.

CHAPTER IV

RIVER TERRACE AND DELTA CHARACTERISTICS

4.1 Introduction

The following discussion focuses on various aspects of the Athabasca River valley geomorphology, distinguishing terrace morphologies and associated alluvial stratigraphies. Terrace sets are defined in terms of the continuity of given terrace surfaces along the valley and the relative elevation of each terrace surface above the present river level. The analysis of terrace alluvium centers on the variations of the mean grain size diameter, sedimentary structures and lithologies of the gravels comprising the terrace sets.

The two major concerns of this study were as follows. First, the morphological expression of the alluvial terraces in the area should provide one base for interpreting the valley history. Prior to the field data collection it was postulated that:

1. terrace development in the upstream sector of the valley resulted mainly from late Quaternary changes in river sediment/discharge relationships, determined largely by activities of the related Cordilleran glacier system,

2. alternating phases of aggradation and degradation throughout the remainder of the river valley were directly influenced by fluctuations of local base levels. Glacial

meltwater, impounded in front of the Laurentide ice sheet, formed temporary base levels to which the Athabasca River graded and regraded as lakes grew and then dissipated, and

3. distinct terrace sets should grade downvalley to limiting elevations of related, now relict, proglacial lake deltas.

A second, more tenuous, basis of interpretation lies in the nature of the terrace alluvium. Differences in grain size, sorting and the primary sedimentary structures of the terrace alluvium attest to the general flow conditions which existed in the Athabasca River during periods of aggradation. As well, the lithologies of the gravels comprising the terrace alluvium are indicative of the general source areas from which the alluvial materials were derived. The detailed mapping of terrace surfaces and sampling of alluvium were somewhat limited by time available, two canoeing accidents and the large field area involved, but the data collected allow first-order comparisons of the major terrace units.

4.2 Methodology

The distribution of terrace remnants initially mapped from aerial photographs was field checked and modified by cut-line traverses and altimetric determinations. The elevations of remnant terrace treads above the present Athabasca River channel were determined and the position of each altimetric transect plotted on the base map, along with the elevation and

width of the river floodplain and valley top elevations.

Where available exposures permitted, the upper limits of bedrock or Pleistocene deposits were noted. Overlying alluvial sequences were measured and described. Major clast lithologies were noted and roundness characteristics assessed qualitatively. The alluvial exposures were photographed for grid-by-number grain size analysis. To obtain accurately representative data on alluvial materials observed in the field large samples must be removed for later analysis, or much time spent on repetitive field measurements. The remoteness and size of the area to be covered in this study prevented use of those two procedures. Rather, a photographic grid-by-number sampling technique was employed. This technique was successfully tested by Bramm (1977) in the context of sampling terrace gravels of selected Alberta rivers. Briefly, the grid-by-number grain size technique allows the approximation of median and mean grain size diameters, d_{sp50} and $\overline{d}_{sp50}$ respectively, from photographs. In the field a 60 centimeter square grid with cross wires at intervals of 5 centimeters was placed on the gravel exposures to be sampled, and photographed (Plate 4-1). Each gravel clast with an apparent b axis equal to or greater than 8 millimeters (Kellerhals and Bray, 1971; Bramm, 1977), and falling beneath a grid intersection on the photograph, was measured and recorded. If a clast smaller than 8 millimeters was located under a grid intersection, the first suitable clast encountered by moving away from the grid point in any of

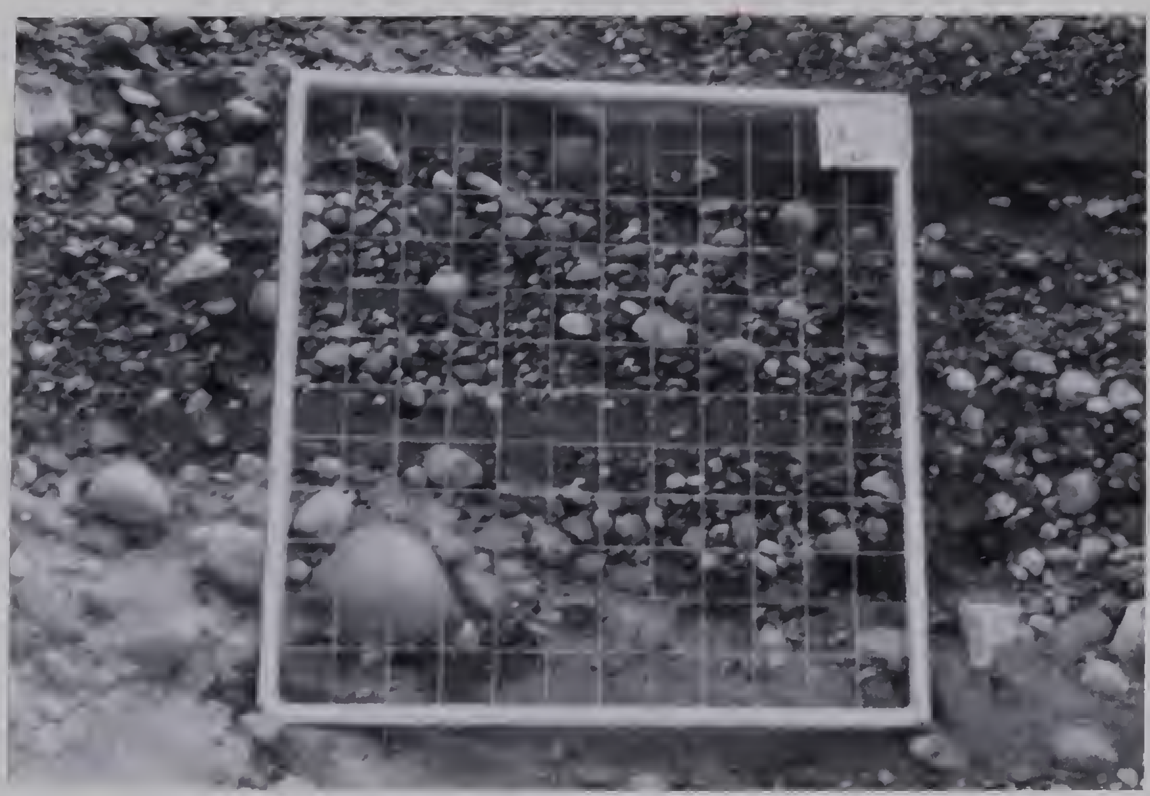


Plate 4-1. Example of grid placement for grid-by-number sampling technique.

the four grid directions, chosen at random, was selected. If two sample points fell on the same clast, the clast was counted twice (Kellerhals and Bray, 1971). For the grid-by-number sample to be considered valid, 50 or more b-axes, measured from one or two photographs, were required.

The apparent major and minor axes (a and b), were measured from the photo and converted to their equivalent phi (Ø) values, using the formula:

$$Ø = \frac{-\log_e(\mathrm{mm})}{\log_e 2} \quad [2]$$

The phi values were then arranged in 0.25 class intervals and cumulative frequency curves for a and b were plotted on arithmetic graph paper. The range of phi values covered by these plots was -8.0 to -3.0. The fiftieth percentile of each plot (a and b) was read from the graphs to give a_{t50} and b_{t50}, the median of each distribution.

The calculations necessary to predict median and mean diameters from grid-by-number photographs were developed by Kellerhals and Bray (1972) and Kellerhals et al., (1975). Kellerhals et al., (1975, p. 83) demonstrated that:

> the mode of the distribution of a- and b- axes were, therefore, computed numerically by cutting ellipsoids with a large number of planes and determining the lengths of the major and minor axes of the elliptic intercepts between the planes and the ellipsoids.

The apparent major axes (a) of the elliptical trace is closely associated with the intermediate axis (B) (Figure 4-1), while

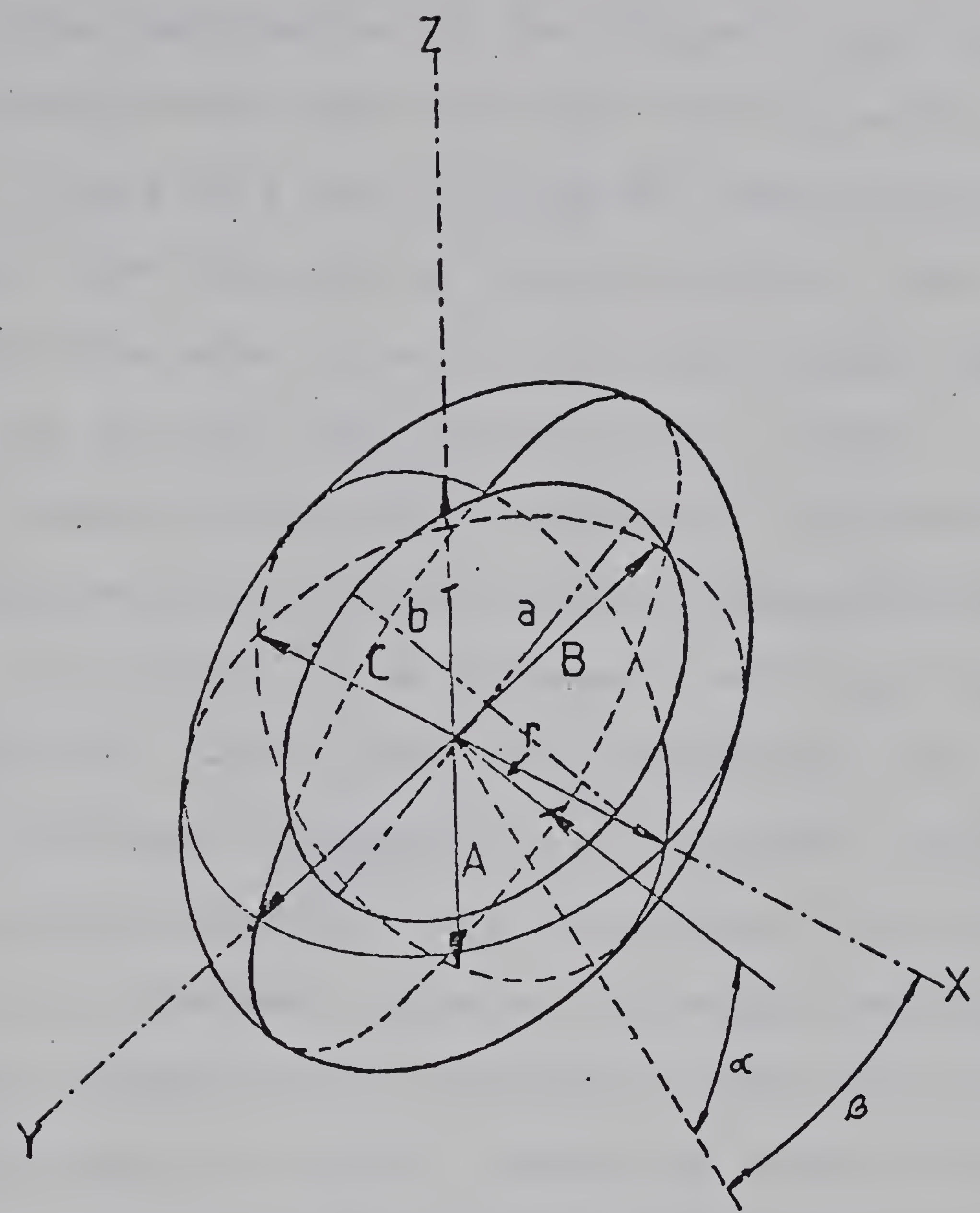


Figure 4-1. Diagram of an ellipsoid intersected by a plane, (after Kellerhals et. al., 1975).

Outline of ellipsoid.
Intersection between ellipsoid and co-ordinate planes.
 Apparent ellipse (intersection between ellipsoid and plane).
Axes of ellipsoid.

Cutting plane is defined by:

α Angle of latitude.
β Angle of longitude.
r Length of normal from the centre of ellipsoid to the cutting plane.
a Apparent major axis.
b Apparent minor axis.

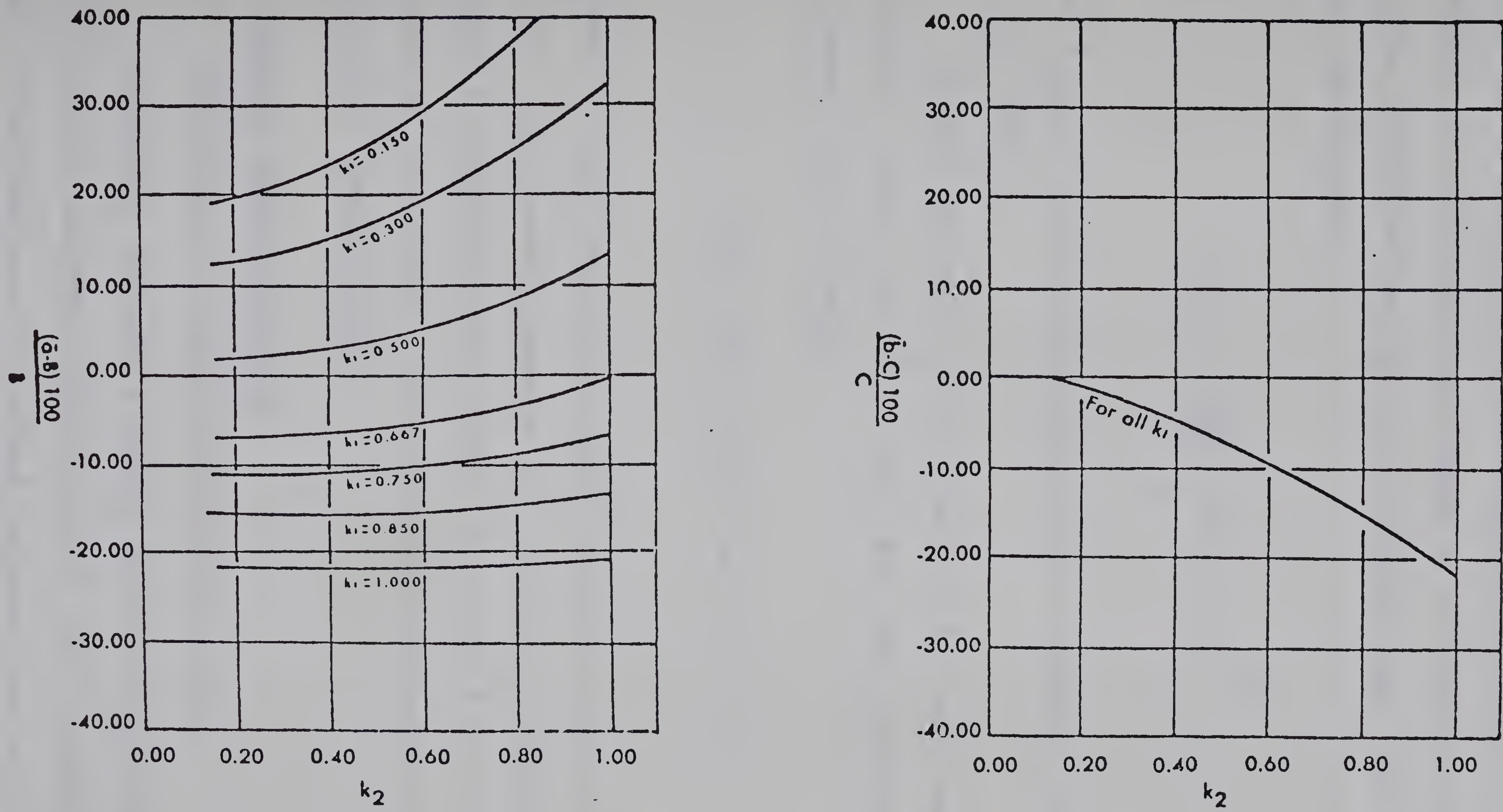
the mode of the distribution of the apparent minor axes (b) is similarily associated with the minor axis (C) (Figure 4-1). Mean values of a and b ($\bar{a}$ and $\bar{b}$), derived from the sample planes were found to be slightly smaller than the corresponding true axes (B and C), for the common range of k values, 0.55 to 0.75 (Kellerhals, et al., 1975).

For cases of non-uniform materials, Kellerhals, et al., (1975) considered this conversion to be theoretically possible but thought it unlikely to be successful with real data. However, Bramm (1977) showed that this relationship did describe non-uniform materials reasonably well and hence could be applied to alluvial terrace gravel sections, the clasts corresponding to the elliptical traces of the model described. The ratio b:a was estimated for the present study from the cumulative distributions of a and b. Using the ratio of median b to median a ($b_{t50}:a_{t50}$), an approximate average k_2 may be obtained (Figure 4-2).

From Figure 4-3 k_2 is used to obtain corresponding values of Y, a correction factor for each sample. This factor is dependent on the ellipsoidal shape and is used in the prediction of median C axis. Hence,

$$C_{p50} = (1.0 - Y)b_{t50} \qquad [3]$$

where C_{p50} is the predicted median C axis of the ellipsoid.



Figures 4-2 and 4-3. Graphs used to obtain Y values for corresponding calculated values of k_2, (after Kellerhals et. al., 1975).

Thus the value of b_{t50} is increased by a certain percentage depending on the elliptical shape of the sample median. Using C_{p50} the predicted median sieve diameter may then be calculated using the formula:

$$d_{sp50} = \frac{C_{p50}}{2K_2}\left[2(1 + k_2^2)\right]^{1/2} \qquad [4]$$

where d_{sp50} is predicted median sieve diameter (Kellerhals, et. al., 1975).

Predicted mean sieve diameters for each sample were calculated using $\bar{a}_t$ and $\bar{b}_t$, which had been calculated from the Inman (1952) formula:

$$\bar{a}_t = \frac{a_{16} + a_{50} + a_{84}}{3} \text{ (similarly for b)} \qquad [5]$$

A summary of the procedures used to obtain the predicted median and mean grain size diameters is contained in Appendix B. Appendix C summarizes the results obtained from the data analysis of 10 samples, denoting predicted median and mean grain sizes, respectively.

4.3 Terrace Characteristics

4.3.1 Morphology and Generalized Stratigraphy

Numerous, paired, alluvial terrace surfaces and occasional unpaired terraces occupy much of the Athabasca River valley. The dominance of paired terraces was evident from

their nearly coincidental elevations and continuity downvalley. However, while consistent relative elevations and continuity are the basic criteria for terrace correlation Frye and Leonard (1954) cautioned that, in order to prevent misleading and possibly incorrect results, the stratigraphy of terrace deposits must also be evaluated. Aspects of the Athabasca River terrace stratigraphy are therefore discussed in Section 4.3.2.

Figure 4-4 shows the location of areas mapped in Figures 4-5 to 4-10 which in turn depict the areal distribution of terrace surfaces in the valley. The locations of alluvial exposures detailed in Appendix A are indicated on the maps. Figure 4-11 illustrates the generalized vertical and downvalley dimensions of the remnant alluvial terrace surfaces, bordering the Athabasca River valley, between Hinton and Whitecourt. Cross-valley profiles illustrate the width and vertical dimensions of various terrace surfaces throughout the study area. Figure 4-4 indicates the position of the valley cross-profiles shown in Figures 4-12 to 4-13 and the positions at which Plates 4-2 to 4-7 were taken. The relative thicknesses of the surficial deposits indicated on the cross-sections are mainly estimates. The composite diagram of Figure 4-14 illustrates a variety of cross-valley profiles typical of the study sector.

For many of the individual valley cross profiles only two major, paired terrace surfaces, with occasional unpaired surfaces, are evident. The assignment of individual terrace renmants to a specific order of the main paired terrace units

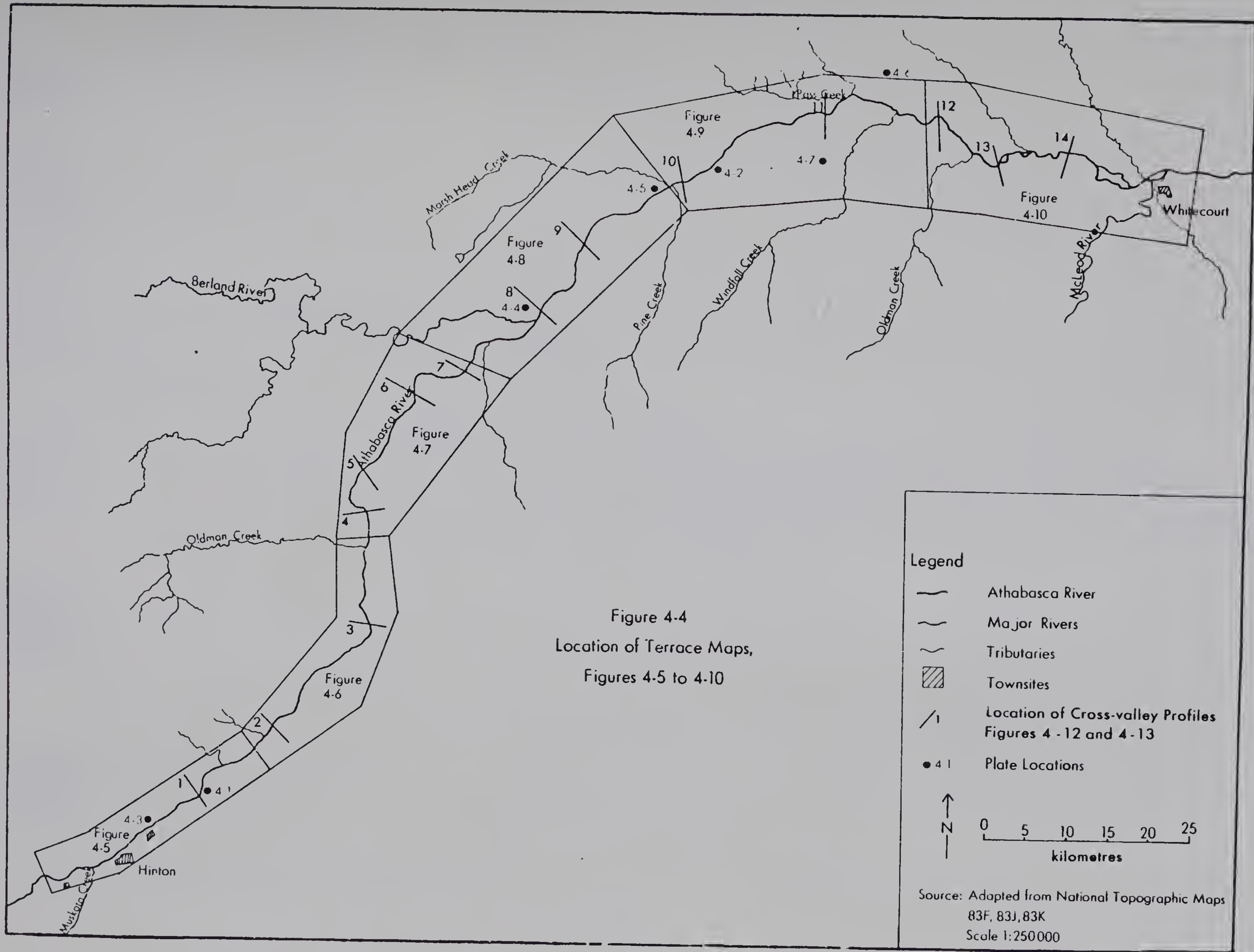


Figure 4-4
Location of Terrace Maps,
Figures 4-5 to 4-10

LEGEND

UNIT	MORPHOLOGICAL DESCRIPTION	TERRACE MATERIAL
	Terrace One (T_1)	Coarse textured, poorly sorted, subangular to subrounded gravels, overlain by thin discontinuous beds of coarse to medium grained sands and subrounded to rounded pebbles. Gravel clasts are primarily quartzites, limestones and dolomites.
	Terrace Two (Hinton) (T_{2H})	Moderately sorted sand and gravel deposits, with interbedded laminated sand lenses. Fining upward cycles occur within the gravel units. Gravels are rounded to well rounded, primarily quartzites, limestones, dolomites and sandstones.
	Terrace Two (T_2)	Poorly sorted, rounded to well rounded gravels in a medium to coarse sand matrix. Little stratification of the gravel units is evident, and only relatively minor, stratified, sand lenses occur. Gravels consist primarily of quartzites, limestones and sandstones.
	Terrace Three (T_3)	Poorly sorted, rounded to subrounded gravels in a medium to coarse sand matrix. Gravels consist primarily of limestones and quartzites with minor traces of igneous clasts in the downstream sector.
	Terrace Four (T_4)	Poorly sorted, rounded to subrounded gravels in a medium to coarse sand matrix. Little evidence of overbank fines exists. Gravels consist primarily of limestones and quartzites, and some igneous clasts.
	Present Floodplain	Poorly sorted, rounded to well rounded gravels in a coarse to medium sand matrix.

* 3-31 Section Locations (See Appendix A)

Break in Slope

Geomorphic Boundary; Defined, Inferred

3500 Contour (feet) (1 foot=.304 meters)

C.I. = 250 feet

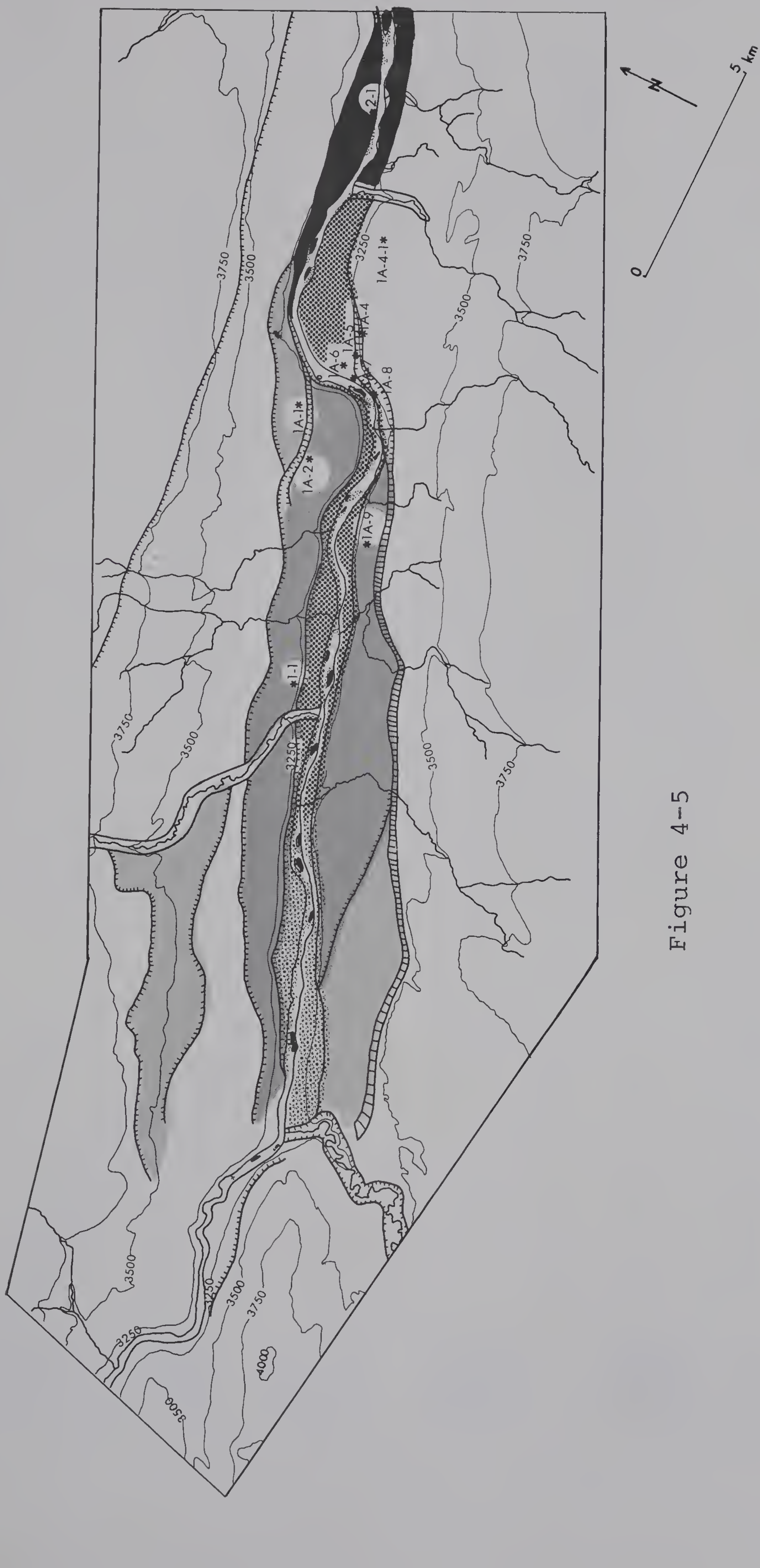

2-1
1A-4-1
1A-4
1A-5
1A-6
1A-8
1A-1
1A-2
1A-9
1-1
3250
3500
3750
4000
N
0
5 km

Figure 4-5

LEGEND

UNIT	MORPHOLOGICAL DESCRIPTION	TERRACE MATERIAL
	Terrace One (T_1)	Coarse textured, poorly sorted, subangular to subrounded gravels, overlain by thin discontinuous beds of coarse to medium grained sands and subrounded to rounded pebbles. Gravel clasts are primarily quartzites, limestones and dolomites.
	Terrace Two (Hinton) (T_{2H})	Moderately sorted sand and gravel deposits, with interbedded laminated sand lenses. Fining upward cycles occur within the gravel units. Gravels are rounded to well rounded, primarily quartzites, limestones, dolomites and sandstones.
	Terrace Two (T_2)	Poorly sorted, rounded to well rounded gravels in a medium to coarse sand matrix. Little stratification of the gravel units is evident, and only relatively minor, stratified, sand lenses occur. Gravels consist primarily of quartzites, limestones and sandstones.
	Terrace Three (T_3)	Poorly sorted, rounded to subrounded gravels in a medium to coarse sand matrix. Gravels consist primarily of limestones and quartzites with minor traces of igneous clasts in the downstream sector.
	Terrace Four (T_4)	Poorly sorted, rounded to subrounded gravels in a medium to coarse sand matrix. Little evidence of overbank fines exists. Gravels consist primarily of limestones and quartzites, and some igneous clasts.
	Present Floodplain	Poorly sorted, rounded to well rounded gravels in a coarse to medium sand matrix.

* 3-31 Section Locations (See Appendix A)
Break in Slope

Geomorphic Boundary; Defined, Inferred
3500 Contour (feet) (1 foot=.304 meters)
C.I. = 250 feet


N
0
5 km
3500
3250
3-8
3-9
3500
3250
3500
3250
3-1
3-1
3-2
3-4-1
2-5
2-3
2-4-1
3500
3750
3500
3250
2-6
2-2
3250
3500
3750

Figure 4-6

LEGEND

UNIT	MORPHOLOGICAL DESCRIPTION	TERRACE MATERIAL
	Terrace One (T_1)	Coarse textured, poorly sorted, subangular to subrounded gravels, overlain by thin discontinuous beds of coarse to medium grained sands and subrounded to rounded pebbles. Gravel clasts are primarily quartzites, limestones and dolomites.
	Terrace Two (Hinton) (T_{2H})	Moderately sorted sand and gravel deposits, with interbedded laminated sand lenses. Fining upward cycles occur within the gravel units. Gravels are rounded to well rounded, primarily quartzites, limestones, dolomites and sandstones.
	Terrace Two (T_2)	Poorly sorted, rounded to well rounded gravels in a medium to coarse sand matrix. Little stratification of the gravel units is evident, and only relatively minor, stratified, sand lenses occur. Gravels consist primarily of quartzites, limestones and sandstones.
	Terrace Three (T_3)	Poorly sorted, rounded to subrounded gravels in a medium to coarse sand matrix. Gravels consist primarily of limestones and quartzites with minor traces of igneous clasts in the downstream sector.
	Terrace Four (T_4)	Poorly sorted, rounded to subrounded gravels in a medium to coarse sand matrix. Little evidence of overbank fines exists. Gravels consist primarily of limestones and quartzites, and some igneous clasts.
	Present Floodplain	Poorly sorted, rounded to well rounded gravels in a coarse to medium sand matrix.

* 3-31 Section Locations (See Appendix A)

Break in Slope

Geomorphic Boundary; Defined, Inferred

3500 Contour (feet) (1 foot=.304 meters)

C.I. = 250 feet

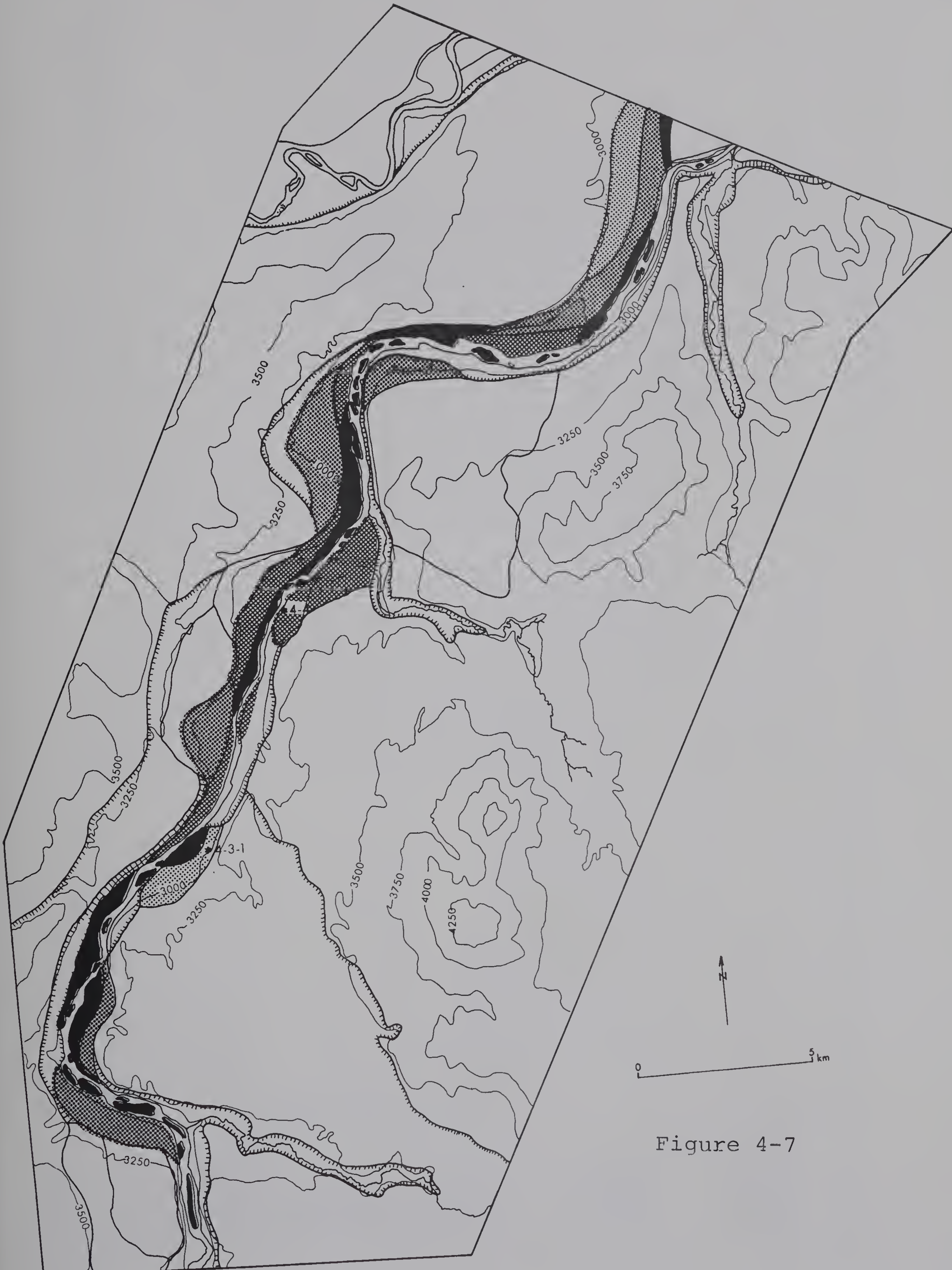

3000
3000
3500
3250
3500
3750
4000
3250
4-
3500
3250
4-3-1
3000
3250
3500
3750
4000
4250
N
0
5 km
3250
3500

Figure 4-7

LEGEND

UNIT	MORPHOLOGICAL DESCRIPTION	TERRACE MATERIAL
	Terrace One (T_1)	Coarse textured, poorly sorted, subangular to subrounded gravels, overlain by thin discontinuous beds of coarse to medium grained sands and sub-rounded to rounded pebbles. Gravel clasts are primarily quartzites, limestones and dolomites.
	Terrace Two (Hinton) (T_{2H})	Moderately sorted sand and gravel deposits, with interbedded laminated sand lenses. Fining upward cycles occur within the gravel units. Gravels are rounded to well rounded, primarily quartzites, limestones, dolomites and sandstones.
	Terrace Two (T_2)	Poorly sorted, rounded to well rounded gravels in a medium to coarse sand matrix. Little stratification of the gravel units is evident, and only relatively minor, stratified, sand lenses occur. Gravels consist primarily of quartzites, limestones and sandstones.
	Terrace Three (T_3)	Poorly sorted, rounded to subrounded gravels in a medium to coarse sand matrix. Gravels consist primarily of limestones and quartzites with minor traces of igneous clasts in the downstream sector.
	Terrace Four (T_4)	Poorly sorted, rounded to subrounded gravels in a medium to coarse sand matrix. Little evidence of overbank fines exists. Gravels consist primarily of limestones and quartzites, and some igneous clasts.
	Present Floodplain	Poorly sorted, rounded to well rounded gravels in a coarse to medium sand matrix.

* 3-31 Section Locations (See Appendix A)

Break in Slope

Geomorphic Boundary; Defined, Inferred

3500 Contour (feet) (1 foot=.304 meters)

C.I. = 250 feet

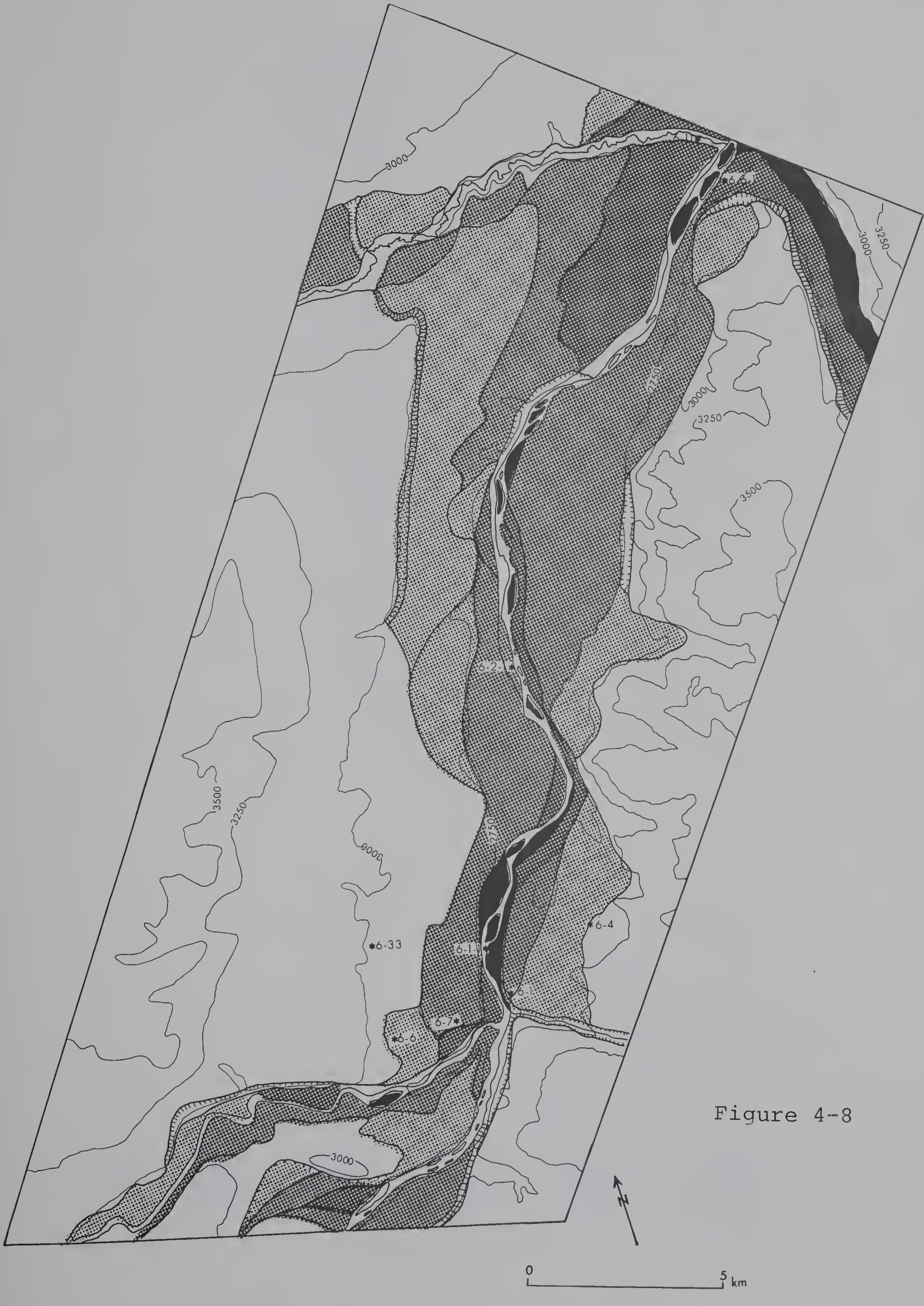


Figure 4-8

LEGEND

UNIT	MORPHOLOGICAL DESCRIPTION	TERRACE MATERIAL
	Terrace One (T_1)	Coarse textured, poorly sorted, subangular to subrounded gravels, overlain by thin discontinuous beds of coarse to medium grained sands and sub-rounded to rounded pebbles. Gravel clasts are primarily quartzites, limestones and dolomites.
	Terrace Two (Hinton) (T_{2H})	Moderately sorted sand and gravel deposits, with interbedded laminated sand lenses. Fining upward cycles occur within the gravel units. Gravels are rounded to well rounded, primarily quartzites, limestones, dolomites and sandstones.
	Terrace Two (T_2)	Poorly sorted, rounded to well rounded gravels in a medium to coarse sand matrix. Little stratification of the gravel units is evident, and only relatively minor, stratified, sand lenses occur. Gravels consist primarily of quartzites, limestones and sandstones.
	Terrace Three (T_3)	Poorly sorted, rounded to subrounded gravels in a medium to coarse sand matrix. Gravels consist primarily of limestones and quartzites with minor traces of igneous clasts in the downstream sector.
	Terrace Four (T_4)	Poorly sorted, rounded to subrounded gravels in a medium to coarse sand matrix. Little evidence of overbank fines exists. Gravels consist primarily of limestones and quartzites, and some igneous clasts.
	Present Floodplain	Poorly sorted, rounded to well rounded gravels in a coarse to medium sand matrix.

* 3-31 Section Locations (See Appendix A)

Break in Slope

Geomorphic Boundary; Defined, Inferred

3500 Contour (feet) (1 foot=.304 meters)

C.I. = 250 feet

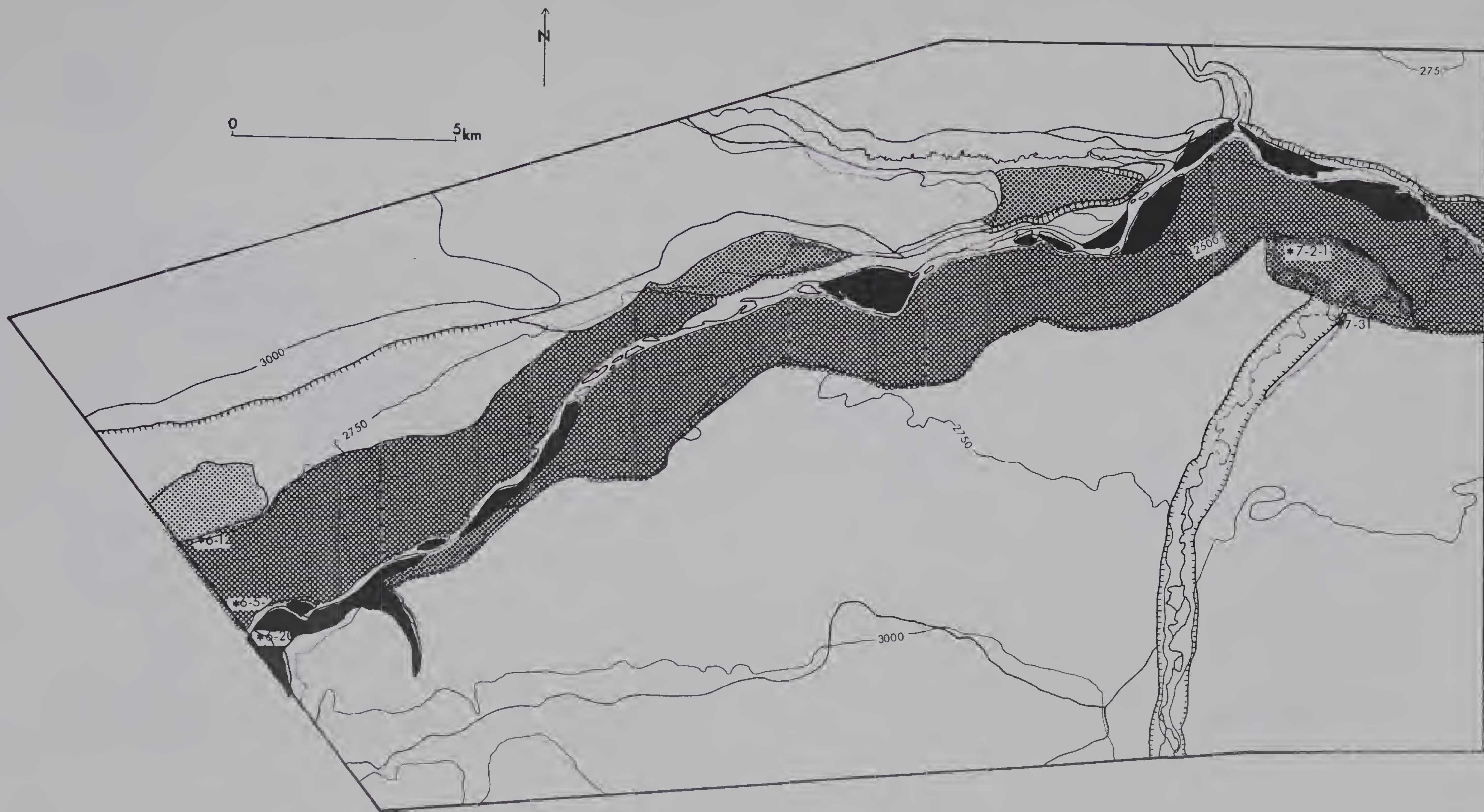


Figure 4-9

LEGEND

UNIT	MORPHOLOGICAL DESCRIPTION	TERRACE MATERIAL
	Terrace One (T_1)	Coarse textured, poorly sorted, subangular to subrounded gravels, overlain by thin discontinuous beds of coarse to medium grained sands and subrounded to rounded pebbles. Gravel clasts are primarily quartzites, limestones and dolomites.
	Terrace Two (Hinton) (T_{2H})	Moderately sorted sand and gravel deposits, with interbedded laminated sand lenses. Fining upward cycles occur within the gravel units. Gravels are rounded to well rounded, primarily quartzites, limestones, dolomites and sandstones.
	Terrace Two (T_2)	Poorly sorted, rounded to well rounded gravels in a medium to coarse sand matrix. Little stratification of the gravel units is evident, and only relatively minor, stratified, sand lenses occur. Gravels consist primarily of quartzites, limestones and sandstones.
	Terrace Three (T_3)	Poorly sorted, rounded to subrounded gravels in a medium to coarse sand matrix. Gravels consist primarily of limestones and quartzites with minor traces of igneous clasts in the downstream sector.
	Terrace Four (T_4)	Poorly sorted, rounded to subrounded gravels in a medium to coarse sand matrix. Little evidence of overbank fines exists. Gravels consist primarily of limestones and quartzites, and some igneous clasts.
	Present Floodplain	Poorly sorted, rounded to well rounded gravels in a coarse to medium sand matrix.

* 3-31 Section Locations (See Appendix A)

Break in Slope

Geomorphic Boundary; Defined, Inferred

3500 Contour (feet) (1 foot=.304 meters)

C.I. = 250 feet

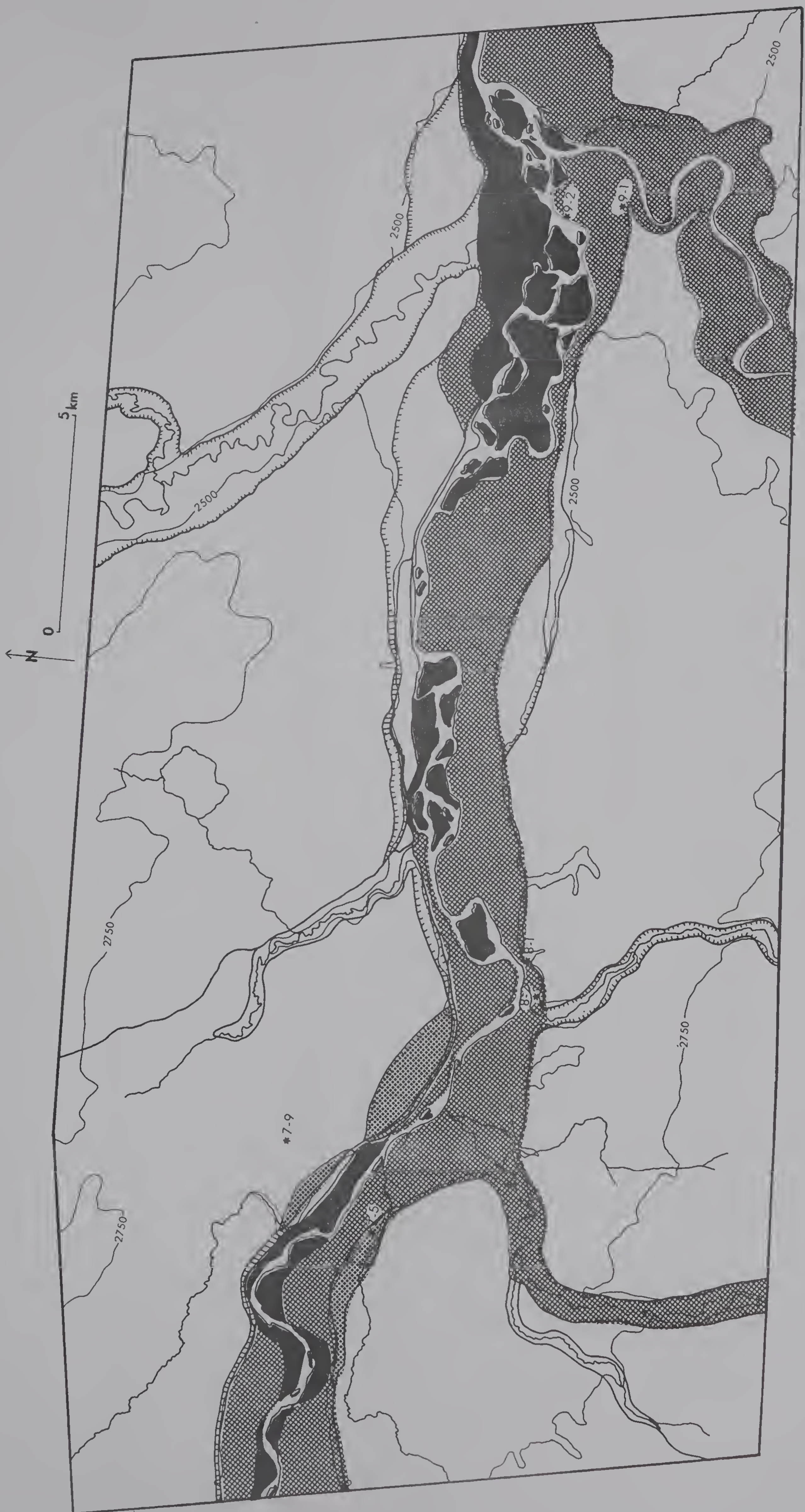


Figure 4-10

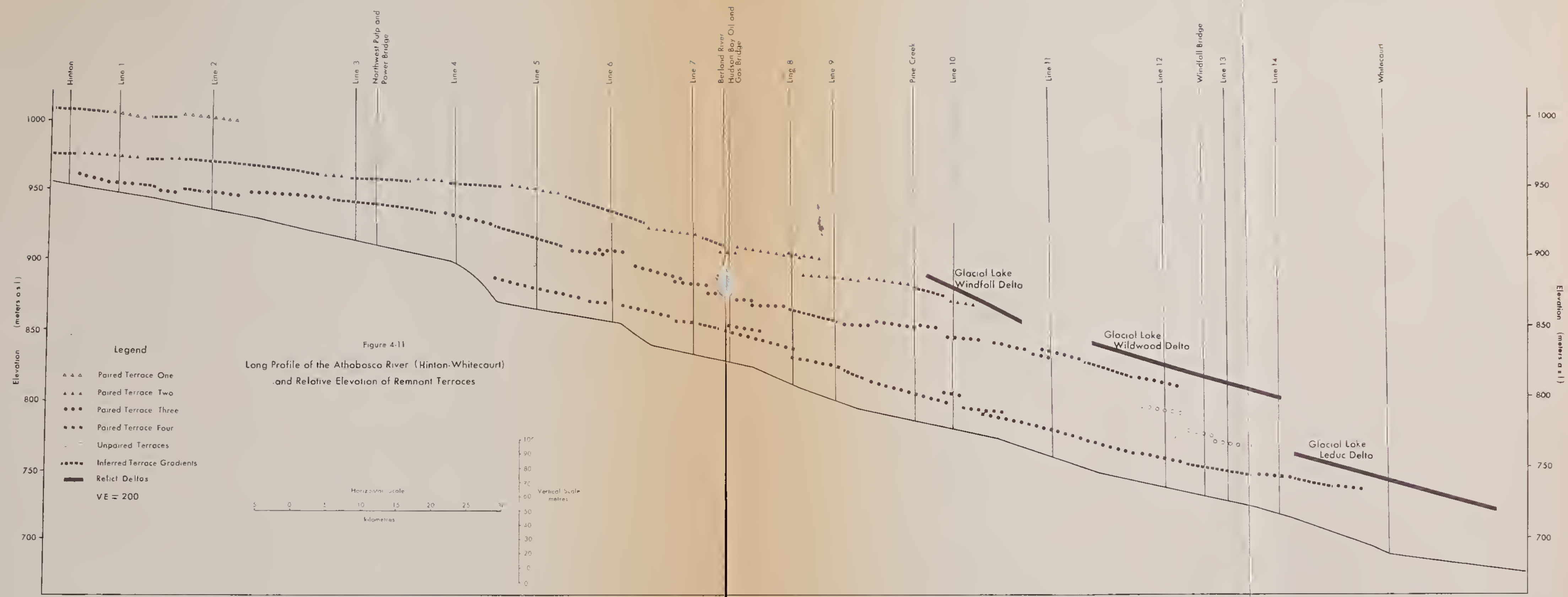


Figure 4-11

Long Profile of the Athabasca River (Hinton-Whitecourt) and Relative Elevation of Remnant Terraces

LEGEND

UNIT	SURFICIAL DEPOSITS
■ ■ ■	Till
● ● ●	Terrace Sand and Gravels
o o o	Outwash Sand and Gravels
△ △ △	Relict Deltaic Sands
□ □ □	Bedrock
▲ ▲ ▲	Colluvium
⊗⊗⊗⊗⊗	Peat/Organics

T1 = Paired Terrace One

T2 = Paired Terrace Two

T3 = Paired Terrace Three

T4 = Paired Terrace Four

FP = Present Valley Floodplain

(V.E. = 25X)

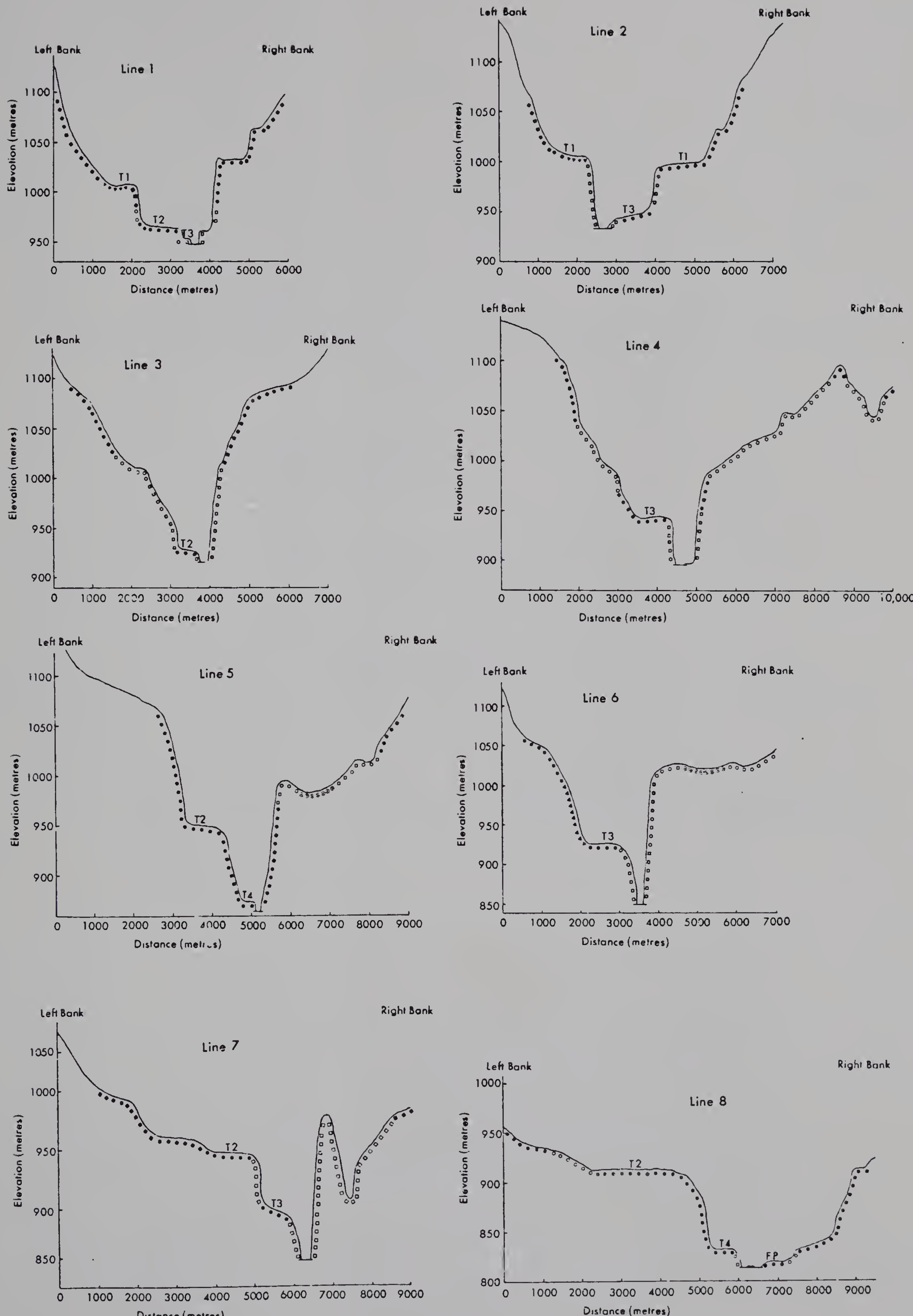


Figure 4 -12

Representative valley cross-sections.

LEGEND

UNIT	SURFICIAL DEPOSITS
■ ■ ■	Till
● ● ●	Terrace Sand and Gravels
○ ○ ○	Outwash Sand and Gravels
△ △ △	Relict Deltaic Sands
□ □ □	Bedrock
▲ ▲ ▲	Colluvium
⊗⊗⊗⊗⊗	Peat/Organics

T1 = Paired Terrace One

T2 = Paired Terrace Two

T3 = Paired Terrace Three

T4 = Paired Terrace Four

FP = Present Valley Floodplain

(V.E. = 25X)

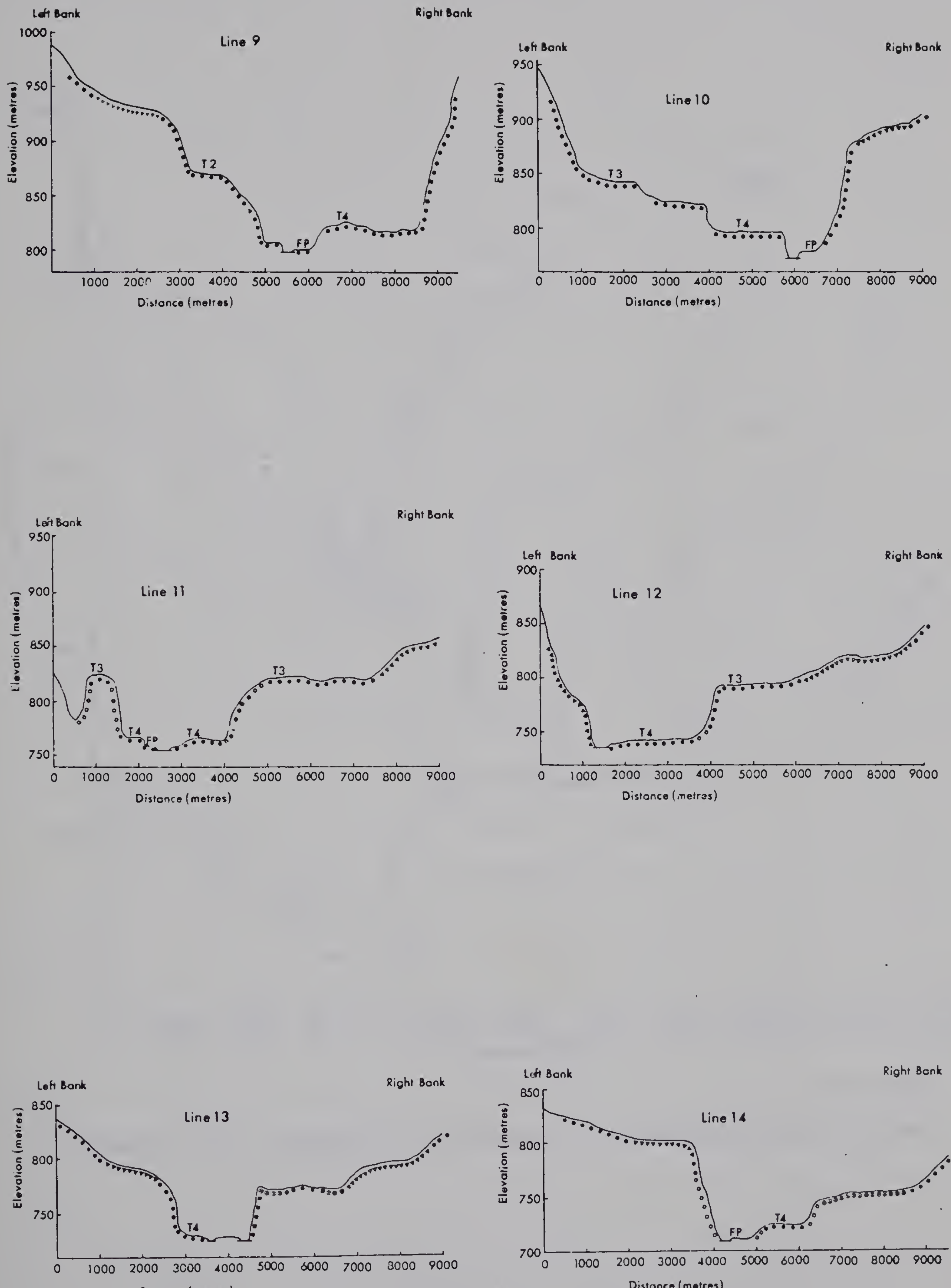
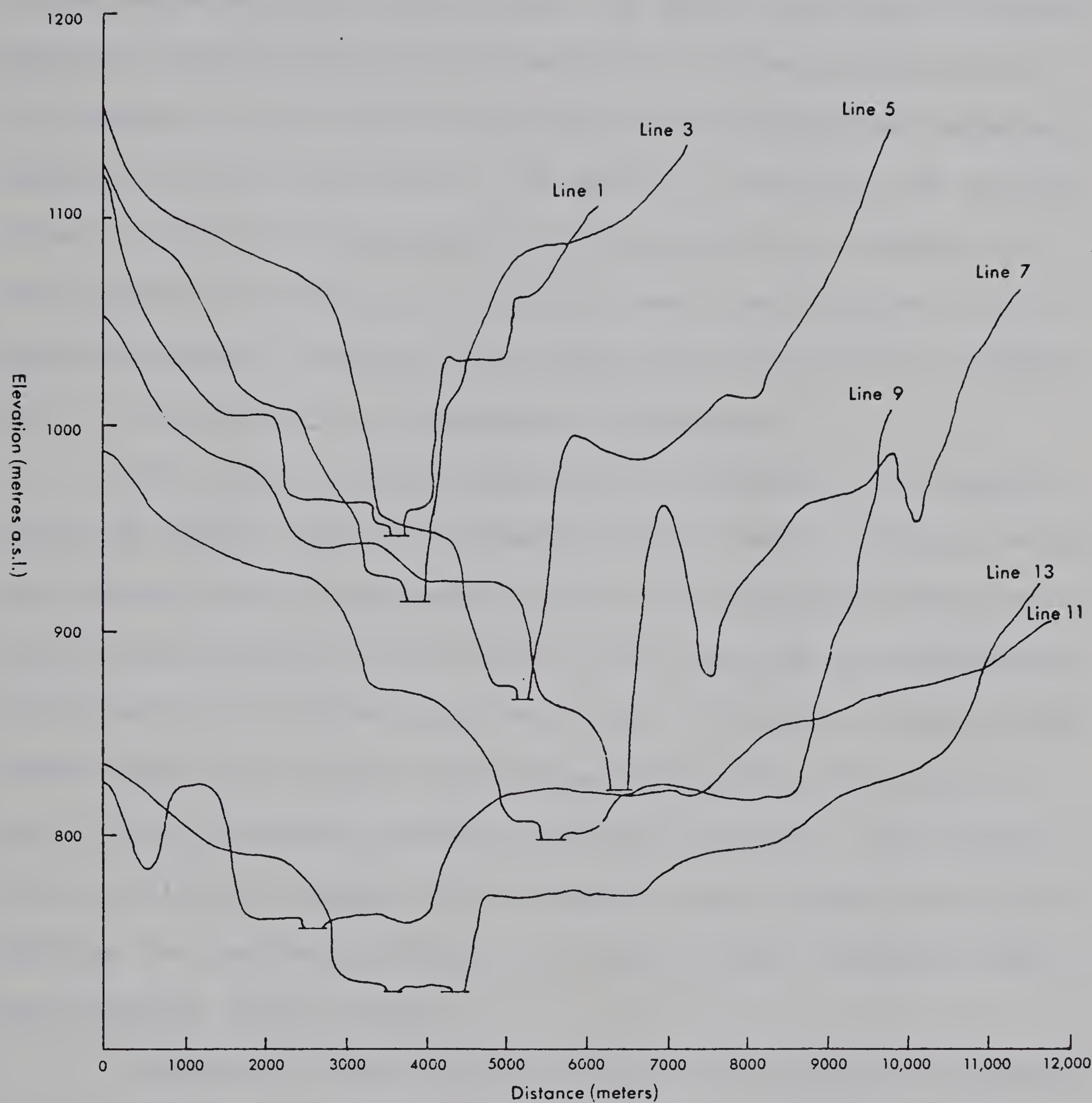


Figure 4-13

Representative valley cross-sections.



V.E.=25X

Figure 4-14. Composite diagram of cross-valley profiles.

was determined by the relative elevation of the alluvial surfaces above the present river channel and the continuity of the surfaces along the valley. At some locations isolated, unpaired surfaces were also identified. From Figures 4-5 to 4-11, however, four distinct paired units and several unpaired surfaces may be identified. For ease of reference the paired terrace units are designated T_1, T_2, T_3 and T_4, ranging from the highest and oldest (T_1) to the lowest and youngest (T_4). Unpaired terrace remnants were identified and plotted accordingly, along with the contemporary floodplain.

The highest paired terrace, T_1, surfaces lie approximately 60 meters above the present river channel (Figure 4-11). This terrace unit is marked by only a few isolated remnants in the upstream sector of the study reach area and it terminates approximately 20 kilometers downvalley of Hinton. Stene (1966) showed that this terrace extends up-valley from the Hinton area. The T_1 terrace treads are very irregular. At certain sites colluvium derived from steeper, upper slopes partially obscures the terrace surfaces; on other treads numerous abandoned channel scars occur.

Alluvium of the highest terrace unit comprises approximately 5 meters of gravels, overlying Pleistocene till deposits. The gravels are generally coarse textured, poorly sorted, subangular to subrounded, and are overlain by thin discontinuous beds of coarse to medium grained sands and subrounded to rounded pebbles. Gravel clasts are composed primarily of

quartzites, limestones, dolomites and sandstones of Cordilleran origin. Little variation in either clast composition or grain size distributions was evident for examples of this terrace unit. The thin, upper layer of gravels and sands may have resulted from the re-working of alluvium by migrating braided channels.

Remnants of terrace unit T_2 are distributed from well upstream of Hinton, 125 kilometers downvalley to the Pine Creek confluence, at which point the T_2 terrace becomes indistinguishable. Terrace surfaces located downvalley of the Berland River mouth are more frequent and extensive than are related treads upstream. The T_2 unit long profile is slightly convex-up in nature (Figure 4-11). Remnant surfaces, situated approximately 30 meters above the present channel near Hinton, extend to 80 meters above the channel in the vicinity of Nosehill Creek (Figure 4-12, #7), and to approximately 90 meters near the Pine Creek confluence. This terrace unit is thus divergent from the present channel in a downstream direction. Terrace surfaces of this unit are frequently marked by abandoned channel scars. Several terrace surfaces are partially mantled by colluvium, derived from the valley sides (Figures 4-12 and 4-13), and others have been considerably truncated by tributary streams. The majority of the T_2 surfaces are well defined though.

The T_2 alluvium consists of 3 to 5 meters of variably sorted gravels, overlying bedrock or till deposits. The

contacts between the bedrock and alluvium are very distinct (Plate 4-2). In the vicinity of Hinton the terrace alluvium consists of a series of moderately sorted sand and gravel deposits, with interbedded laminated sand lenses. Fining upward cycles occur within the gravel units (Plate 4-3). Gravels are rounded to well rounded, primarily quartzites, but include some limestone and sandstone clasts. The alluvium has a medium to coarse sand matrix. Downvalley of Hinton, to the Berland River confluence, the T_2 deposits consist of poorly sorted, rounded to well rounded gravels, ranging from 10 to 50 centimeters in diameter, in a medium to coarse sand matrix. There is little apparent stratification of the gravel units and only relatively minor, stratified, sand lenses. Gravels consist primarily of quartzites, with some local sandstones and limestones included (Plate 4-4). Minor coal fragments, were also observed. The alluvial stratigraphy and grain size characteristics of the terrace deposits remain relatively consistent throughout the remainder of the terrace sequence. However, downvalley of the Berland River confluence minor quantities of Canadian Shield igneous and metamorphic clasts are included in the alluvium.

Remnants of terrace unit T_3 are the most evenly distributed and continuous members within the study area. The unit extends from well up valley of Hinton to the Windfall Creek confluence, 180 kilometers downvalley. No T_3 terrace surfaces were found downstream of that confluence. The valley profile

Plate 4-2. Distinct contact between terrace alluvium and underlying bedrock.

Plate 4-3. Fining-upward of alluvial gravels, T_2.

Plate 4-4. Illustrated T_2 section 6-7.

of the terrace unit is very similar to that of the T_2 distribution in that it is convex in nature. Terrace surfaces diverge in relative elevation above the channel from approximately 5 meters near Hinton, to 40 meters near the Berland River confluence and up to 60 meters in the Windfall Creek area. The T_3 surfaces are in most cases very well defined, although some treads have been partially obscured by colluvium (Figures 4-12 and 4-13), or truncated by tributary stream incision (Figure 4-13). The most extensive surfaces are nearly horizontal, displaying numerous abandoned channel scars. The absence of overbank fines throughout much of the terrace set, abandoned channel scars on the terrace treads, and the nearly horizontal profile of the terrace surfaces all tend to suggest that braided channel deposition (see Donjek type, Chapter I, Section 1.3) existed prior to channel incision. When channel incision did begin it was relatively rapid and continuous. The inferred rapid channel incision resulted in the abandonment of former braided channels.

Alluvial deposits making up the T_3 unit usually overlie bedrock with a sharp contact (Plate 4-5) and vary in thickness from 2 to 5 meters. The alluvium consists of poorly sorted, rounded to subrounded gravels in a medium to coarse sand matrix. Gravels range from 10 to 20 centimeters in diameter (Plate 4-6). Little change in either the clast sizes or matrix composition was evident throughout the study reach but gravel lithologies did vary slightly. Upstream of the

Plate 4-5. Contact between T_3 alluvium and underlying bedrock.

Plate 4-6. Illustrated T_3 section 7-30.

Berland River confluence the gravels consist primarily of quartzites, with some limestones, sandstones and minor coal fragments of local origin. Downstream of the Berland River area the gravels consist of local sandstones, quartzites, limestones and very minor quantities of Canadian Shield metamorphic and igneous clasts. The frequency of metamorphic and igneous clasts in this alluvium appears, though, to be considerably less than for the T_2 alluvium described earlier.

The youngest, lowest, paired terrace unit (T_4) tends to closely parallel the present river long profile. Originating approximately 60 kilometers downstream of Hinton the terrace continues downstream of Whitecourt. The elevations of the remnant surfaces making up the terrace set vary between 5 meters above the present channel in the vicinity of Oldman Creek, to approximately 15 meters above the present channel in the Whitecourt area (Figure 4-11). The T_4 terrace treads are the most extensive and well preserved of all the Athabasca River valley terrace units. Remnant surfaces in the upstream portion of the study area are not large but downvalley of the Berland River confluence the Athabasca River valley widens, the T_4 surfaces increase markedly in size. The treads are approximately horizontal and are marked with channel scars. Some of the terrace remnants have developed in conjunction with channel splitting around major islands (Figure 4-13).

The T_4 deposits comprise 1 to 5 meters of alluvium overlying bedrock, the intervening contact being very distinct.

The thickness of the alluvium fluctuates partly with variations of the underlying bedrock surface. Alluvium of the T_4 unit consists of poorly sorted, rounded to well rounded gravels in a coarse to medium sand matrix. Gravels range from about 15 to 25 centimeters in diameter and there is little evidence of overbank fines or fining upward cycles in the gravels. Occasional, laminated, sand lenses were observed but only at a few localities. The gravels consist primarily of quartzites and limestones, with a greater percentage of sandstones than was observed in the T_2 and T_3 alluvium. Some metamorphic and igneous clasts were found in the terrace gravels, downstream of the Berland River confluence, but only at a few localities.

Several unpaired terrace remnants were identified throughout the study reach. While it is recognized that the terrace remnants which deviate slightly from the generalized terrace unit elevations (Figure 4-11) may be unpaired, only those surfaces at markedly anomalous elevations have been designated as unpaired. Unpaired terrace surfaces are most evident at elevations between the T_3 and T_4 terrace sets. Two such surfaces were identified in the upstream sector of the valley, with three being identified at various locations downstream of the Berland River confluence. Their deposits comprise a thin veneer of poorly sorted sands and gravels overlying bedrock, and in this respect the remnants resemble strath terraces. It is most probable that, as in many river valleys, the unpaired terraces simply reflect fortuitously preserved segments

of former floodplains developed during relatively accentuated phases of river downcutting.

In summary two main generalizations may be stated. First, paired terraces in the study area seldom occur on both valley walls at any one cross profile location. It is only from a composite long profile of the terrace treads, in relation to the present channel, that the relative concentrations into discrete paired terrace units becomes clear. Second, many of the remnant surfaces are marked by abandoned channel scars. This fact suggests that at times of terrace formation channel splitting may have been more pronounced than at the present.

4.3.2 Grain Size Analysis

Initial field logging of alluvial terrace stratigraphies showed that in most instances, except for minor variations in clast size and composition, sediments of the three lowest paired aggradational sequences, T_2, T_3 and T_4, were rather similar. A lack of exposures of the highest terrace, T_1, alluvium prevented analysis of this sequence. However, the T_2 terrace unit, in the vicinity of Hinton, did show some variations of alluvial stratigraphy with respect to the remainder of the terrace suite. Therefore further attention was focused on these apparent differences.

Descriptions of all alluvial sections used in this analysis are presented in Appendix A, outlining the major

elements of the depositional sequences. In general the T_2 alluvium is composed of rounded to subrounded gravels and exhibits a series of fining upward cycles throughout the sections. Interbedded, contorted, sand lenses occur in all sections studied in the Hinton area. Further downvalley, bedded gravels are less pronounced, with only occasional contorted sand beds being present. A thin layer of coarse sands overlies the gravel units in some cases. The alluvium of T_3 terraces consists mainly of rounded gravels in a coarse to medium sand matrix. There is little evidence of gravel bedding but a few sections exhibit a slight fining upward sequence. No sections of this sequence were found upstream of the Berland River confluence. Materials of the T_4 unit are very similar to those of the T_3 members and consist of well rounded gravels in a sandy matrix. Fining upward sorting and gravel bedding are very poorly developed except in some cases where thin lenses of laminated sands overlie the gravels.

Given these gross similarities of the sediments, and their stratigraphies, the gravel characteristics showed that subtle differences might be distinguished. A more precise illustration of possible differences in the alluvial gravels of the major terrace units is given by the frequency distributions of the values determined by grid-by-number, grain size calculations (Figure 4-15). To ascertain whether differences in the mean grain sizes of the various terrace deposits were significant the Student t-test was applied (Hammond and

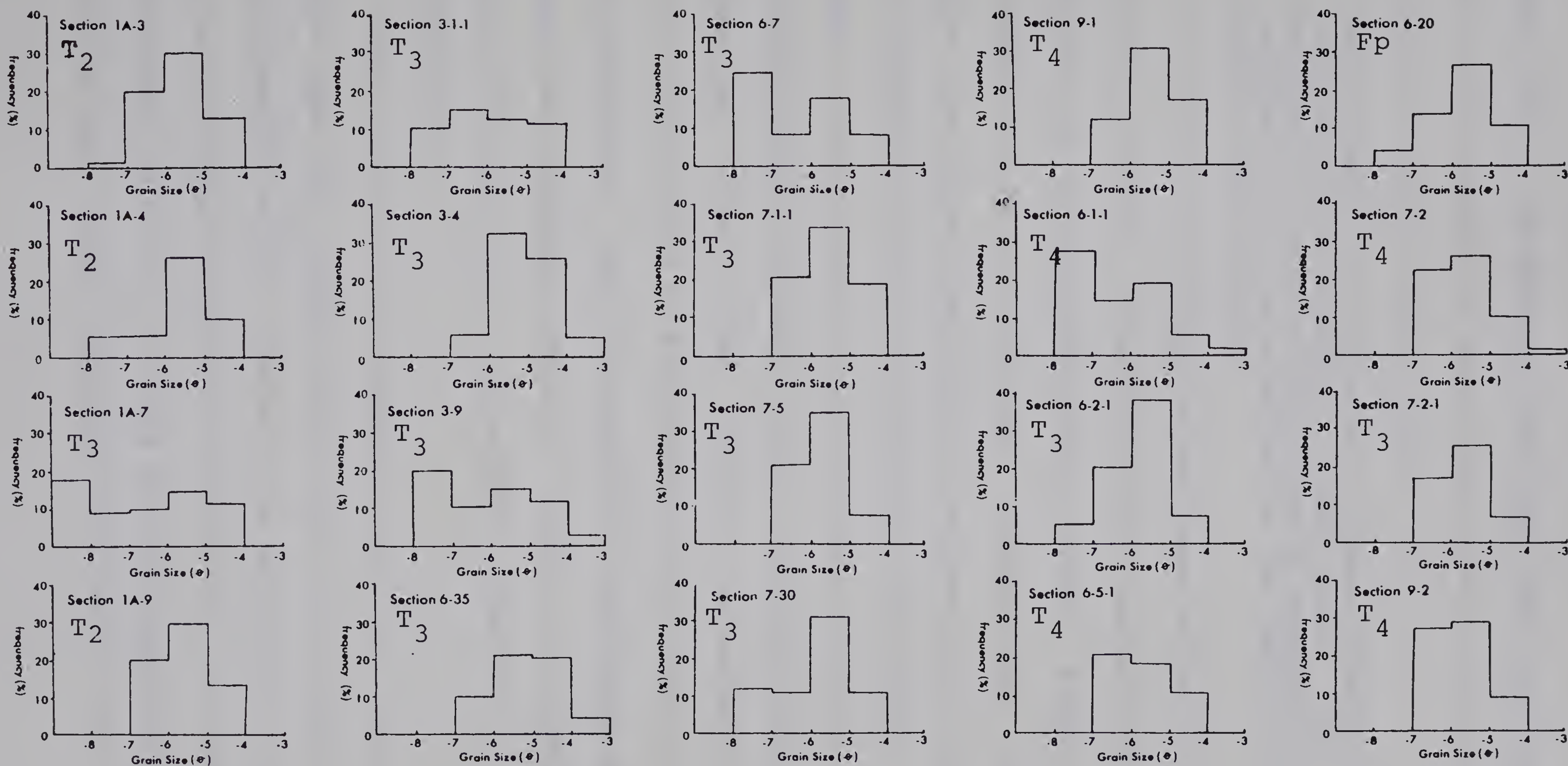


Figure 4-15. Frequency distribution of alluvial gravels determined by grid-by-number grain size calculations.

McCullagh, 1974). In applying the test, the apparent b axis was used as the measure of mean grain size. Kellerhals and Bray, (1971, p. 1169), stated that,

> Once a sample is collected a linear dimension of grain size has to be assigned to each grain. The intermediate or b axis of the grain or measures closely related to it; are accepted as the definition of grain size by most authors.

For strict statistical comparisons of grain size distributions, however, several criteria should be met before the analysis can be considered wholly valid. First, populations sampled are to be independent of each other. As discussed earlier, the three lowest paired terrace units occur at markedly differing elevations throughout the valley. The variations of their vertical distribution indicate that they are of significantly different ages, and hence may be considered independent of each other. Second, the method of sampling should have been stratified to eliminate any bias in the sampling procedure. While the sampling procedures employed here cannot be considered strictly random, as suggested by Cole and King (1970), any bias in the sample collection by the author was unconscious.

The validity of the Student t-test is based on the assumption that the standard deviations of the sample populations are approximately equal. Whether this assumption is valid or or not can be tested using the variance ratio or F-test. If the calculated value of F is less than the appropriate value indicated by the F tables (Murdoch and Barnes, 1970) then the

sample variance may be considered identical (Cole and King, 1968). Table 4-1 shows that the sample variances are approximately equal and thus application of the Student t-test to the sample means was carried out. On the basis of the general sediment characteristics noted earlier it was postulated (null hypotheses, Ho) that:

1. the mean grain sizes of alluvium for the major terrace units are not significantly different, and

2. the mean grain size of T_2 terrace alluvium near Hinton (T_{2H}), does not differ significantly from that of T_2 alluvium further downvalley (T_2).

The results of the t-test are outlined in Table 4-2. The test was carried out at the 5 percent significance level with the degrees of freedom as calculated in the Student t-test equation (Appendix D). Briefly, the results indicate that there is probably no significant difference between the average grain sizes of the T_3 and T_4 unit alluvium, but that the other combinations tested probably have significantly differing mean grain sizes. The implications of the similarities and differences of grain sizes shown here are discussed further in Chapter V.

4.4 Delta Characteristics

Earlier work by St-Onge (1972) showed that, during the retreat of the Laurentide ice sheet from the study area, water from a variety of sources was impounded in parts of the

TABLE 4-1

F-TEST

Units Tested	F_c	F	Test Results*
T_4/T_3	4.53	1.09	Accept Ho
T_4/T_2	4.53	1.31	Accept Ho
T_3/T_2	6.39	1.09	Accept Ho
T_{2H}/T_2	9.28	1.16	Accept Ho

*Tested at the 5% level of significance

TABLE 4-2

T-TEST

Units Tested	t_c	t	Test Results*
T_4/T_3	1.81	1.02	Accept Ho
T_4/T_2	1.83	3.98	Reject Ho
T_3/T_2	1.89	2.87	Reject Ho
T_{2H}/T_2	1.94	6.32	Reject Ho

*Tested to the 5% level of significance

Athabasca River valley west of the Laurentide ice front. The resulting series of proglacial lakes probably grew and dissipated relatively rapidly as the ice front continued to waste northeastward. In this manner new outlet channels for these lakes rapidly evolved, as did the continued extension and trenching of the main Athabasca River valley. During the growth of each new lake, however, the lower reaches of the Athabasca River immediately west of each successive lake margin, experienced temporary net aggradation rather than degradation. The most conspicuous products of this aggradation are numerous delta complexes, identified by St-Onge (1972), plus their upvalley extensions in the form of partial alluvial valley fills.

Three large delta complexes occur within the study area. Each of the three large delta complexes determined here may infact not be representative of a single deltaic unit but rather a series of smaller deltaic units formed at various levels during the let-down period of each proglacial lake sequence. However, in the field,these smaller individual deltaic units could not be distinguish with any certainty, so reference to only the three primary deltaic units is made. Related, ancient lake shorelines and good exposures of lacustrine sediments are comparatively rare. Thus it was necessary to estimate former proglacial lake levels from the elevations of these relict delta surfaces. The areal extent of each proglacial lake associated with the three main

deltas could not be accurately checked in this study. Therefore, to maintain consistency, the mapped proglacial lake limits and lake names associated with each overall delta complex have been derived from the work of St-Onge (1972).

The oldest delta complex, D_1 (Figure 3-5), extends from east of the Berland River confluence downstream to the approximate vicinity of the Pass Creek confluence. The delta was constructed from materials laid down by the Athabasca River and Smoky River/Marsh Head Creek spillway channel as they emptied into the western end of Glacial Lake Windfall, between 870 and 840 meters above sea level (Figure 4-11). Glacial Lake Windfall drained to the south through an outlet channel 10 kilometers southwest of Whitecourt (Figure 3-9). The delta consists of approximately 3 meters of medium to fine sands, overlying till. The fine sands are fairly well consolidated with minor clay/silt lenses overlain by medium sands. The sand units are cross-stratified and parallel bedded (Plate 3-16).

Meltwater flowing through the Pass Creek spillway into the Athabasca River valley and Glacial Lake Wildwood contributed to the formation of the middle delta complex, D_2, entending from Windfall Creek to the Oldman Creek confluence (Figure 3-6). Deposits making up this deltaic unit vary from 2 to 4 meters in thickness. The sediments consist of medium to fine sands, with minor clay lenses and scattered coal fragments. The sand units display large scale, high angle, planar cross-

strata. The middle delta complex varies in elevation from 840 meters at its westernmost limit to 820 meters in the east (Figure 4-11). The areal extent, elevation range, gradient and sedimentary composition of the middle delta complex are consistent with that of delta D_1.

The lower delta complex, D_3, which begins in the vicinity of the Oldman Creek confluence, at an elevation of 750 meters, terminated downvalley at an elevation of 720 meters, east of Whitecourt (Figure 4-11 and 3-7). Water continued to flow through the Athabasca River and Pass Creek spillway channel into Glacial Lake Leduc, and was responsible for the deposition of the deltaic materials (Figure 3-11). Sediments making up the lowest delta complex are very similar to those of the upper two units. Approximately 3 to 5 meters of medium to fine grained sands overly localized bedrock outcrops and lacustrine sediments. The sands, while not highly consolidated, do exhibit a series of planar cross-bedded and horizontally bedded structures. Minor gravel and coal lenses were also identified.

With the lowering and eventual disappearence of each glacial lake, as a result of the continued retreat of the Laurentide ice margin, these tracts of deltaic sands were fully exposed and were entrenched by the Athabasca River. Before vegetation succession could be established on the relict surfaces winds were able to deflate the deltaic deposits, creating several dune fields throughout the area. These dune fields are well preserved to-day (Plate 4-7).

Plate 4-7. Aerial photograph of remnant dune field. (Scale 1:2640) (Plate location shown on Figure 4-4).

CHAPTER V

INTERPRETATION AND CONCLUSIONS

5.1 Introduction

Alluvial terraces of the Athabasca River valley, between Hinton and Whitecourt, developed mainly during the late Quaternary retreat of the associated Cordilleran and Laurentide ice masses. This episode may have begun approximately 12,000 to 14,000 years B.P. (St-Onge, 1972; Luckman, 1977; Kvill, 1979) or possibly earlier. The factors most likely to have been responsible for the alternating cut and fill valley development during this period include:

1. postglacial tilting of the land surface as a result of isostatic rebound,
2. localized obstruction of the river drainage by ice-damming of the Laurentide ice sheet,
3. Cordilleran glacier advance and retreat,
4. related changes in the sediment/discharge relationships of the Athabasca River.

Little is known about the specific amounts or effects of postglacial isostacy or tectonics in this area. However, three arguments against isostacy as a major factor controlling terrace development may be suggested:

1. generally the thickness of the Laurentide ice sheet would have increased northeastward from the study area, located close to the ice sheet margin at its maximum extent. Thus isostatic depression and rebound in the study area were probably of minor significance.

2. the occurrence of proglacial lake and deltaic deposits in association with the major paired terrace units suggests that glacier ice was still in close proximity during their formation.

3. it is difficult to envisage how three or four episodes of accentuated isostatic rebound, leading to separate degradational stages, might be triggered in the study area.

Although isostatic movements probably made some contributions to the cut and fill history of the river valley these were almost certainly of minor significance. The comparative importance of the three remaining controlling factors in this area is by no means certain because of their obvious interdependence, and the very primitive state of the known geochronology. Given these substantial constraints the following interpretation of valley evolution is presented below. Figures 3-1 through 4-11 complement the following discussion. As well, the idealized valley profile, Figure 5-1, should be consulted while reviewing the following interpretation.

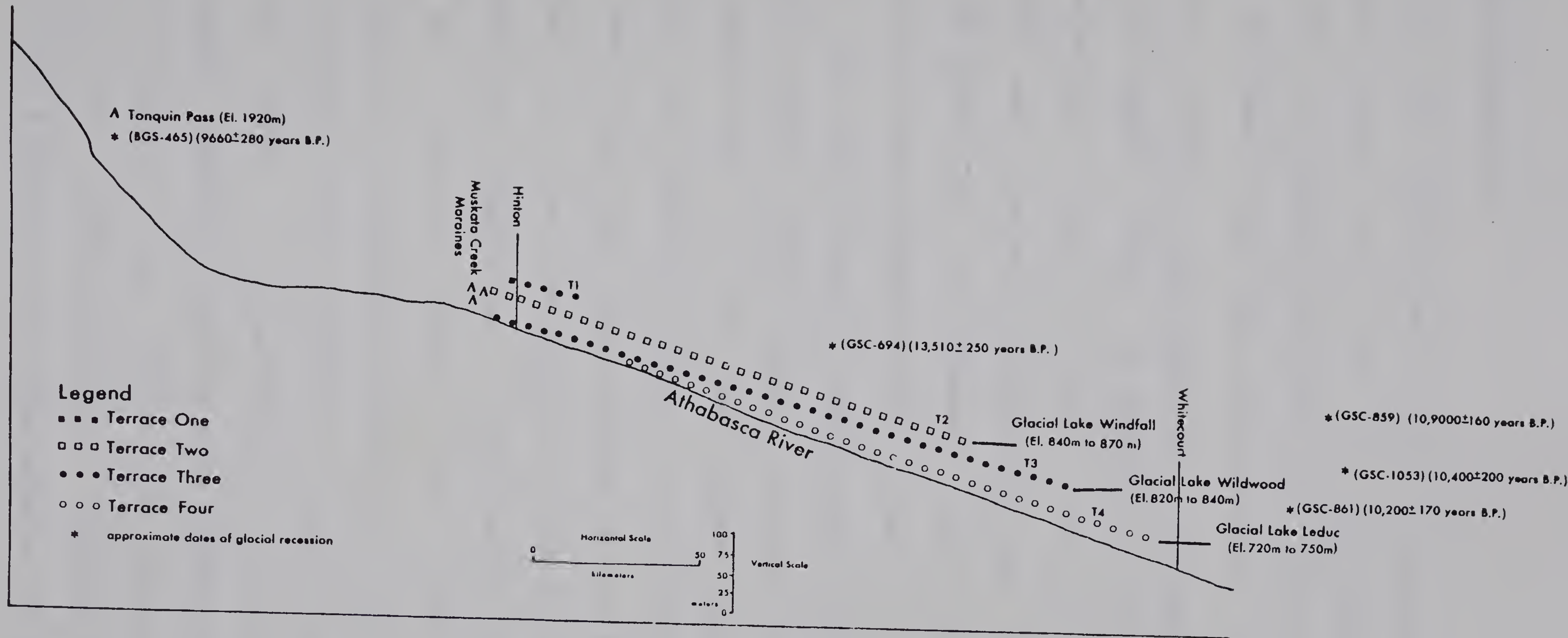


Figure 5-1. Idealized Athabasca Valley profile. (V.E.=25X)

5.2 Interpretation

The last Cordilleran advance to directly influence the western section of the area, the Obed Advance of Roed (1968, 1975), has been correlated with the Late Portage Mountain Advance in the Peace River Valley, which had ended by at least 11,600 ± 1,000 years B.P. (I-2244A, Rutter, 1972; Luckman and Osborn, 1977). In addition to this evidence from Portage Mountain, a date of 11,225 ± 125 years B.P. (S-1705; Kvill, 1979) was obtained from Fairfax Lake, along the eastern edge of the foothills, 50 kilometers south of Hinton. Together these limiting dates suggest that recession of the Cordilleran glaciers, from Lake Wisconsin maximum positions at or outside the mountain front, probably began before 11,500 years B.P.

Tonquin Pass is a low pass (1920m) between tributaries of the Fraser and Athabasca Rivers, 27 kilometers southwest of Jasper. Small esker and kame deposits on the floor of the pass indicate that the final phases of deglaciation in this area involved ice stagnation. A radiocarbon date of 9,600 ± 280 years B.P. (BGS-465, Luckman et. al., 1977), from a basal peat bog found in a depression between the ice stagnation features of Tonquin Pass, provides a limiting date for deglaciation in this region. The peat deposits overlie nonglacial deposits which, together with the time needed for peat formation, suggested to Luckman et al., (1977) that deglaciation had occurred prior to 10,000 years B.P. Further morphological evidence of ice stagnation are kame moraine terraces flanking

the Athabasca Valley near Jasper (Luckman and Osborn, 1977), the Muskata Creek kame moraine (Roed,1975) and Obed kame complex (Figure 3-3). This evidence indicates that the late Quaternary deglaciation of the upper Athabasca River valley may have been rapid, dominated by ice stagnation and down wasting.

To the east dates of 13,510 ± 250 years B.P. (GSC-694) and 12,190 ± 350 years B.P. (GSC-508) (St-Onge, 1972) from deltaic sands of the Smoky River valley provide a minimum age for the recession of the Laurentide glacier from the area. Further dates of 10,900 ± 160 years B.P. (GSC-859), 10,400 ± 200 years B.P. (GSC-1053) and 10,200 ± 170 years B.P. (GSC-861) (St-Onge, 1972) from speciments found in sands/silt deosits east of Whitecourt (Figure 3-12) also suggest that Quaternary deglaciation of the eastern extent of the study area was very rapid, occurring at least 13,000 years B.P. While some evidence for dominant downwastage of Cordilleran ice does exist, there is little evidence to suggest that Laurentide deglaciation to the east resulted from predominant downwasting. Rather, the occurrence of vast tracts of glaciolacustrine and related sediments throughout the valley suggest that glacial recession apparently resulted from pronounced frontal retreat. Together, from the limiting dates available, it appears likely that late Quaternary deglaciation of the entire study area was apparently time-synchronous and had been accomplished by at least 10,000 years B.P.

Initially, Pleistocene advances of the Cordilleran glacier from the west, and the Laurentide glacier from the east, partly infilled the Athabasca Valley system. The two ice masses are considered to have undergone multiple advance and retreat stages (see Section 3.2) but it was not until the final wastage and retreat phases of the Cordilleran and Laurentide glaciers, beginning approximately 12,000 to 14,000 years B.P., that the Athabasca River began to etch out its present day valley.

It seems likely that at first the Laurentide ice sheet receded from its frontal maximum position near the Oldman Creek confluence (Figure 3-8) and stabilized in the vicinity of the Nosehill Creek confluence. During this period, or perhaps earlier, the Cordilleran glacier had wasted upstream of Hinton; the increase in meltwater discharge associated with downwasting served to incise the previously deposited Pleistocene sediments, re-establishing a new Athabasca River course. Drainage to the northeast was still impeded by the Laurentide ice lobe occupying the valley near Nosehill Creek. However, meltwater incision was still able to initiate the present drainage course through the Hinton area to Nosehill Creek. Lacustrine sediments in the vicinity of Oldman Creek (Figure 3-8) suggest that meltwater was ponded in front of the Laurentide ice at this time, forcing drainage from the Athabasca River to flow to the southeast through a low level, outlet spillway (Figure 3-1) occupied today by Sundance Creek.

A possible minor re-advance of the Cordilleran glacier in the main Athabasca Valley (Brule Lake moraine of Heusser, 1956), (Luckman and Osborn, 1979), may have altered channel sediment/discharge relationships in the Hinton area which until this time had been actively eroding the channel. This period of glacial advance served to reduce meltwater discharge while at the same time reducing the distance over which glaciofluvial materials were transported. The ensuing decrease in discharge, coupled with increased sediment loads (in relation to the previous discharge), resulted in the outwash aggradation of the valley in the Hinton area. The reduction in discharge was also reflected in the distance of potential sediment transport and hence the relatively minor downstream extent of the related, T_1, aggradational sequence (Figure 4-11). The analysis of the alluvial materials comprising the highest terrace, T_1, (see Section 4.3.1) records a swift flowing but rapidly fluctuating stream flow, possibly associated with glacial advance, the uppermost surface layer of these outwash gravels being formed while the ice fluctuated close to its maximum extent.

Renewed disintegration and wastage of the Cordilleran glacier, and the continued retreat of the Laurentide ice front, followed this aggradational period. With the renewed downwasting of Cordilleran ice meltwater discharge was again increased. Together with the increase in channel gradient resulting from

the continued retreat of the Laurentide glacier exposing ground at lower elevations, the river incised the previously deposited outwash sediments, creating the highest terrace, T_1. Materials derived from the wasting Cordilleran glacier, and eroded out of the previously deposited outwash sediments, were carried downvalley and deposited near the Berland River confluence where they mixed with outwash materials from the nearby Laurentide ice mass to form the Berland River outwash (Figure 3-5). The absence of an outlet spillway in this area, and the relatively sandy nature of the Berland outwash deposits (illustrated section 6-33) indicate that relatively minor ponding conditions persisted in the Athabasca Valley along the Laurentide ice front. Partial drainage of the valley system probably continued through the Sundance Creek and minor high level spillways. Eventually further recession of the Laurentide glacier continued to increase the channel gradient and incision of the Berland River outwash sediments and underlying bedrock ensued.

Continued downwasting of the Cordilleran glacier, and the stabilization of the Laurentide ice lobe occupying the valley to the east, promoted proglacial lake formation along the ice front near Pass Creek (Figure 3-9). This stimulated a second valley fill period. The characteristics of the terrace gravels at Hinton (T_{2H}) (see Section 3.3.2, illustrated section 1-1) are two fold. First, the lower gravel section of this unit consists of large, poorly sorted gravel clasts, indicative of a swift flowing but rapidly fluctuating stream

flow. The initial increase in channel gradient brought about by the retreat of the Laurentide glacier, and the continued melting of the Cordilleran glacier to the west, could explain the high but variable discharge velocities through the valley The rates of meltwater discharge fluctuated during periods of high and low Cordilleran glacier melt. Outwash sediments derived from the downwasting Cordilleran glacier, and from the erosion of the previous deposits, were transported down-valley, the larger clasts being deposited first, forming the lower depositional sequence of this valley train outwash. Smaller outwash clasts were carried further downvalley and deposited, as discharge velocities were further reduced as meltwater entered Glacial Lake Windfall (Figure 3-9). The upper unit of the Hinton terrace, T_{2H}, is more typical of the remainder of the T_2 alluvium, consisting of smaller, poorly sorted gravels, with a coarse sand matrix, apparently deposited under more uniform flow conditions. Two possible reasons may be offered to explain this change in deposition. First, stabilization of the channel gradient to the new base level, established by Glacial Lake Windfall, would reduce discharge velocities and hence their sediment transport capabilities. The progressive upstream deposition of sediments would ensue. This process of deposition was probably responsible for the deposition of T_2 alluvium further downvalley. The sediments making up this valley fill were derived from the re-working of Pleistocene tills, previously deposited outwash materials, and

Cordilleran outwash sediments, transported downvalley. However, it seems unlikely that the influence of a temporary base level control, 120 kilometers downvalley, would exert sufficient influence to alter channel hydraulics, particularly channel discharge velocities, in the Hinton area. A second, more likely explanation for this change in terrace alluvium composition may relate to the continued downwasting of the Cordilleran glacier. As this ice mass disintegrated small ponds and lakes would have developed in low lying depressions and hollows. Meltwater discharge through these ponded areas would remain relatively constant but the sediment transport potential of these meltwaters would be greatly reduced. Flow velocities were temporarily reduced, through the lacustrine sediment "sinks", allowing only transport and downstream deposition of smaller clast fractions.

Continued recession of the Laurentide ice sheet to the northeast, and a gradual lowering of Glacial Lake Windfall, followed this second aggradational period. A lake system was maintained by the influx of meltwater to the valley during the retreat of the Laurentide ice sheet, but at successively lower elevations. Alluvial fines, carried to the lake margins, built the gently sloping and extensive delta sequence, D_1, (Figure 4-11), associated with this initial lake sequence. The eventual lowering of the temporary base level established by Glacial Lake Windfall, and continued meltwater flow through the valley also led to a second period of channel incision, creating the

T_2 terrace set.

A second period of proglacial lake ponding occurred in the Athabasca Valley near the Windfall Creek confluence, forming Glacial Lake Wildwood (Figure 3-10). With the decrease in elevation and downvalley recession respectively, of Glacial Lake Wildwood, the increase in channel gradient of the Athabasca River had two outcomes—first, the incision of the previous valley fill and underlving bedrock, and second, the eventual divergence of the gradients of T_2 and T_3, and the downvalley extension of T_3 beyond that of T_2 (Figure 4-11). The graded profile of T_3 corresponds to the elevation of the Glacial Lake Wildwood delta, D_2. The eventual stabilization of the channel gradient to the new second, temporary base level led to lateral river migration and erosion of the valley walls upstream of the lake system. The lateral erosion of the previously deposited materials and local bedrock increased sediment loads through the channel. Coupled with the lowering of discharge velocities as the Athabasca River drained into Glacial Lake Wildwood, this promoted a third aggradational fill period, progressively backfilling the valley upstream of lake margins.

It is likely that by this time the Cordilleran glacier was no longer exerting any maior influence, in terms of sediment load, on the channel hydraulics of the study reach. Continued wastage of the ice would have allowed even more extensive ponding of water in low lying, upstream depressions. As well, the ice probably would have already receded beyond Brûle Lake,

the lake acting as a dominant sediment trap for materials being discharged from the remaining Cordilleran ice.

Exposure of a low outlet spillway southeast of Thorsby, as the Laurentide ice lobe continued to recede, led to the rapid lowering of Glacial Lake Wildwood. The increase in relative channel incision by 20 meters, in response to the lowering of the previous temporary base level, resulted in the abandonment of former channels responsible for the deposition of the previous valley fill sequence. This renewed period of incision created T_3.

The Laurentide glacier continued to recede beyond Whitecourt until the ice lobe again stabilized sufficiently long to allow the formation of a third proglacial lake, Glacial Lake Leduc (Figure 3-12). Incision of the Athabasca Valley through the previous alluvial fill and underlying bedrock continued until stabilization of the channel gradient to the new base level was achieved. Lateral migration of the channel followed, eroding previously deposited alluvial and lacustrine sediments, and local bedrock. A low energy discharge environment appears to have prevailed through the reach at this time, as little evidence of high energy depositional facies were observed in the T_4 alluvium (see Section 3.3.2). Stabilization of the channel gradient to the new base level, the increase in sediment load from the lateral erosion of previous alluvial deposits and bedrock and a reduction in discharge velocities, may have led to the progressive backfilling of the lower valley fill upstream of the lake margins (D_3) (Figure 3-12).

Lake recession was gradual as the Laurentide glacier continued to retreat eastward. Eventually a new outlet channel was exposed, and Glacial Lake Leduc, and the Athabasca River, drained to the southeast. Erosion of the alluvial fill to form the lower terrace, T_4, followed with the lowering of the previous base level.

In summary, with the recession of the Pleistocene glaciers from the area, the Athabasca River was able to incise its present course through previous valley fill sediments. Initially, the extent of this valley incision was controlled by fluctuations in sediment/discharge relationships in the channel, and eventually by the temporary base level controls exerted on the channel by the retreating Laurentide glacier. In turn, valley widening and destruction of high terrace units, followed with the stabilization of the channel gradient to new lower base levels. The materials eroded from the valley walls during these periods of lateral incision were utilized for the next valley fill sequence.

5.3 Conclusion

The section of the Athabasca River valley examined in this study is unique in that it affords an opportunity to establish a formal link between the late Quaternary regional activities of Cordilleran and Laurentide glaciers in west-central Alberta. The interpretation of remnant terraces and alluvial stratigraphic sections was used in order to establish a

link between the two glacial systems. The present investigation thus had three main objectives. The first was to establish the pattern and distribution of terrace remnants within the river valley. To this end the extent and elevation of remnant terrace surfaces were surveyed and mapped. General characteristics of alluvial grain size distributions were also examined. The second objective was to assess the prominent factors responsible for terrace formation in both the upstream and downstream valley sectors. The final aim was to integrate this information and present an interpretation of the late Quaternary history of the valley.

Figure 5-1 outlines the variety of terrace relationships which have resulted from variations in upstream and downstream controls on the Athabasca River. Four paired terrace levels are evident within the study reach. The oldest, highest terrace (T_1) is confined to the westernmost sector of the valley. Its related terrace treads lie approximately 60 meters above the present channel at Hinton, and are poorly defined, extending only 28 kilometers into the study area. The remaining three terrace sets are of much greater downvalley extent and are thus of greater consequence to this study. The lower three terrace sets are divergent in nature, with successively lower terraces being more extensive downvalley. Near Hinton, terraces T_2 and T_3 stand approximately 30 and 5 meters above the channel, respectively, but downvalley their relative elevations above the channel increase to meet their limits at

the relict delta complexes, D_1 and D_2. Terrace T_4, which originates approximately 60 kilometers downvalley of Hinton, tends to parallel the present channel, maintaining an approximate elevation of 5 meters above the channel and terminating at the relict delta, D_3.

Tracing the continuity of the terrace remnant surfaces downvalley was not without problems. The narrow, rock walled, V-shaped valley in the upstream sector of the reach severely restricted terrace preservation. Terrace alluvium, which may have been deposited during aggradational periods, was almost completely eradicated during subsequent periods of channel incision, and those surfaces which did survive tend to be narrow and poorly defined. Aside from valley constraints, tributary stream dissection and mass movement modification along the higher terraces also made their identification difficult. Fortunately, however, the lower terrace surfaces were not altered to the same degree. Despite these difficulties the construction of a relative long-valley profile of the terrace sets was possible (Figure 4-11). This profile illustrates the existence of the four main terrace levels, their downvalley continuity, and verifies their largely divergent nature. The divergent pattern of the terrace surfaces may be directly attributed to controls exerted on the Athabasca River by the Cordilleran and Laurentide glaciers. Two fundamental controls were responsible for alluvial terrace development: (1) an upstream control related mainly to high rates of

sediment production from the Cordilleran ice and extraglacial slopes, and (2) a downstream control in the form of base level changes during the episodic retreat of the Laurentide glacier.

The T_1 valley fill period apparently resulted from a minor advance stage of the Cordilleran glacier which decreased the ratio of meltwater discharge to available sediment and lead to partial valley filling. Renewed glacial wastage subsequently increased meltwater discharges allowing incision of the alluvial fill to form T_1.

A grain size change in the stratigraphy of T_{2H} suggests that its upstream control during the second valley fill period must also have changed. The lower, coarser T_{2H} deposits were laid down by meltwaters from the downwasting Cordilleran glacier which achieved high flow and correspondingly high sediment concentrations. Later, a decrease in discharge velocities and upstream sediment supplies, resulted in the deposition of the smaller clasts which comprise the upper T_{2H} alluvial unit.

The primary factor responsible for filling and cutting of the T_2, T_3, and T_4 alluvium, as well as the divergent nature of these terraces, was the downvalley development of Laurentide, proglacial lake systems at various altitudes. Three relict proglacial lake deltas which represent both the former lake margins, and the downvalley termination points of terraces T_2, T_3 and T_4 are the main evidence for the proposed base

level control of terrace development. Each proglacial lake acted as a temporary base level to which the river graded. Once the channel gradient stabilized to this level the predominant river behavior would have been lateral erosion of previous valley fills and local bedrock. In this manner progressive backfilling of alluvium upvalley of the lake deltas occurred. Further retreat of the Laurentide ice episodically lowered the proglacial lakes, allowing incision of the previous valley fill to form the terraces.

In terms of general sedimentary characteristics the alluvial sequences of terraces T_2, T_3 and T_4 provide some broad indicators of the hydraulic conditions which prevailed during their periods of deposition. The average textures of alluvium in terraces T_3 and T_4 were found to be very similar (Table 4-2). Both units overly bedrock and are dominated by moderately sorted, well rounded clasts averaging approximately 4.0 centimeters in diameter , with a coarse sand matrix. This alluvial sequence is relatively consistent throughout both terrace units suggesting that these two terrace fills were deposited under similar and comparatively uniform flow conditions. Terrace T_2, however, exhibits a wider range of sediment textures and structures. Near Hinton the T_{2H} alluvium consists of rounded, bedded gravels, averaging 5.0 centimeters in diameter in the lower part of the unit and 2.0 centimeters in diameter in the upper part of the terrace. Interbedded sand lenses also occur.

Placed in context with the earlier work of Stene (1966), Roed (1968, 1975), St-Onge (1972) and Luckman and Osborn (1977) the characteristics of the Athabasca River terraces may be interpreted as follows. The downwasting and recession of the Cordilleran Obed glacier, and the retreat of the Laurentide ice mass, began in this area approximately 12,000 to 14,000 years B.P. The oldest terrace development in the western sector of the study area resulted mainly from meltwater discharge/outwash sediment fluctuations, dictated by activities of the Cordilleran ice. In contrast, development of the younger terraces was more directly controlled by the temporary base levels of the proglacial lakes along the receding Laurentide ice front. Of particular interest is the association and downvalley gradation of the Cordilleran controlled T_{2H} formation with the Laurentide, base level controlled, terrace T_2. This terrace reflects best the sought-after link between the two glacial systems. It indicates the approximate ice frontal positions of the two glaciers during one phase of their recession and also that recession of the Cordilleran and Laurentide glaciers from this area probably occurred simultaneously.

Attempts to find datable materials in the terrace alluvium were unsuccessful. However, by relating the minimum limiting dates for various phases of Laurentide retreat (Figure 3-2) (St-Onge, 1972), and the downwasting of the Cordilleran glacier (Luckman and Osborn, 1977) to the interpreted evolution

of this valley the following general conclusions are drawn;

(1) the available dates suggest that west-central Alberta was ice free by at least 10,000 years B.P., but probably earlier,

(2) the limiting dates of 11,600 ± 1000 years B.P. (I-2244A) obtained by Rutter (1972) for the Peace River and 11,225 ± 125 years B.P. (S-1705) from Kvill (1979) for the Brazeau River, immediately south of Hinton, suggest that recession of the Cordilleran glaciers from their late Wisconsin maximum position beyond the mountain front probably began prior to 11,500 years B.P. Further dates of 13,510 ± 230 years B.P. (GSC-694) and 12,190 ± 350 years B.P. (GSC-508) obtained by St-Onge (1972) from the Smoky River valley, immediately north of the Athabasca/Berland River confluence, suggest recession of the Laurentide glacier from its late Wisconsin maximum position began prior to 13,000 years B.P. Together these dates indicate that the deglaciation of this portion of west-central Alberta was relatively rapid, occurring over a period of 2,000 to 3,000 years.

(3) Finally, based on the above limiting dates, and the interpreted associations of terrace remnants with glacial controls in the Athabasca River

valley, recession of the Cordilleran and Laurentide glaciers in this area appears to have occurred simultaneously.

BIBLIOGRAPHY

BIBLIOGRAPHY

Allen, J.R.L. 1964. Studies in fluviatile sedimentation: six cyclothems from the Lower Old Red Sandstone, Anglo-Welsh Basin. Sedimentology, v. 3, pp. 163-168.

Allen, J.R.L. 1965. The sedimentation and palaeogeography of the Old Red Sandstone of Anglesey, North Wales. Proc. Yorks. Geol. Soc., v. 35, pp. 139-185.

Alley, N.F. 1973. Glacial stratigraphy and the limits of the Rocky Mountain and Laurentide ice sheets in southwestern Alberta, Canada. Bull. Can. Pet. Geol., v. 21, pp. 153-177.

Alley, N.F. and Harris, S.A. 1974. Pleistocene glacial lake sequences in the Foothills, southwestern Alberta, Canada. Can. J. Earth Sci., v. 11, pp. 1220-1235.

Boydell, A.N. 1978. Multiple glaciations in the Foothills, Rocky Mountain House area, Alberta. Res. Council Alberta Bull., No. 36.

Bramm, R. 1977. Terrace Gravel Photographic Grid Sampling. M.Sc. thesis, Geog. Dept., Univ. of Alberta, pp. 119

Bray, D.I. 1972. Generalized Regime-type Analysis of Alberta Rivers. Ph.D. thesis, Dept. Civ. Eng., Univ. of Alberta, 232 pp.

Bretz, J.H. 1943. Keewatin end moraines in Alberta, Canada. Bull. Geol. Soc. Am., v. 54, pp. 31-52.

Cole, J.P. and King, C.A.M. 1968. Quantitative Geography. London, Wiley and Sons, 692 pp.

Church, M. and Gilbert, R. 1975. Proglacial Fluvial and Lacustrine Environments, in Glaciofluvial and Glacio-lacustrine Sedimentation. Soc. Econ. Pal. Min., Special Pubn., No. 23.

Collins, G.A. and Swan, A.G. 1955. Glacial geology, St. Anne Area. Res. Council Alberta Rept. 67.

Cotton, C.A. 1950. Classification and Correlation of River Terraces. J. Geomorph., v. 3, pp. 27-37.

Culling, W.E.H. 1957. Equilibrium States in Multicyclic Streams and the Analysis of River Terrace Profiles. J. Geol., v. 651(5), pp. 451-467.

David, P.P. 1964. Surficial Geology and Groundwater Resources of the Prelate Area, 72K Saskatchewan. Ph.D. Thesis, McGill University, 329 pp.

Dury, A.H., (ed.), 1970. Rivers and River Terraces, Praeger Publ., New York, 283 pp.

Flint, R.F. 1957. Glacial and Pleistocene Geology. John Wiley and Sons Inc., New York, 553 pp.

Folk, R.L. and Ward, W.C. 1957. Brazos River Bar, A Study in the Significance of Grain Size Parameters. J. Sed. Petrol., v. 27(1), pp. 3-26.

Frye, J.C. and Leonard, A.R. 1954. Some Problems of Alluvial Terrace Mapping. Am. J. Sci., v. 252, pp. 242-251.

Hammond, P. and McCullagh, P.S. 1974. Quantitative Techniques in Geography. Clarendon Press, Oxford, 318 pp.

Harris, S.A. and Boydell, A.N. 1972. Glacial History of the Bow River and Red Deer River Areas and the Adjacent Foothills, in Slaymaker, O. and McPherson, H.J. (ed.), Mountain Geomorphology. Tantalus Res, Ltd., Vancouver.

Harris, S.A. and Waters, R.R. 1977. Late Quaternary History of Southwest Alberta: A Progress Report. Bull. Can. Pet. Geol., v.251(1), pp. 35-62.

Henderson, E.P. 1959. Surficial Geology of Sturgeon Lake Map Area, Alberta. Geol. Surv. Can. Mem., 303.

Hillman, A.R., Powell, J.H. and Rothwell, R.L. 1978. Hydrometeorology of the Hinton-Edson area, Alberta, 1972-1975. Northern Forest Research Cetre Information Rept. NOR-X-202, 171 pp.

Howard, A.D., Fairbridge, R.W. and Quinn, J.H. 1968. Terraces, Fluvial-Introduction in Encyclopedia of Geomorphology, (ed. R.W. Fairbridge), New York, Reinhold, pp.1117-24.

Heusser, C.J. 1956. Postglacial Environments in the Canadian Rocky Mountains. Ecological Monographs, v. 26., pp. 253.

Inman, D.L. 1952. Measures for Describing the Size Distribution of Sediments. J. Sed. Petrol., v. 22, pp. 125-145.

Irish, E.J.W. 1965. Geology of the Rocky Mountains Foothills Alberta, (between latitudes 53°15' and 54°15') Geol. Surv. Can. Mem. 334, 241 pp.

Jackson, L.E. 1979. New Evidence for the Existence of an Ice Free Corridor in the Rocky Mountain Foothills near Calgary, Alberta, during Late Wisconsin time. Geol. Surv. Can. Paper, 79-1A, pp. 107-111.

Kellerhals, R. and Bray, D.I. 1971. Sampling Procedures for Coarse Fluvial Sediments. J. Hyd. Div. A.S.C.E., v. 97, No. H87, pp. 1165-1180.

Kellerhals, R., Neill, C.R. and Bray, D.I. 1972. Hydraulic and Geomorphic Characteristics of Rivers in Alberta. Res. Council Alberta, River Engineering and Surface Hydrology, Rept. 72-1.

Kellerhals, R., Shaw, J. and Arora, V.K. 1975. On Grain Size From Thin Sections. J. Geol., v. 83, pp. 79-96.

Kvill, D.R. 1979. Glacial Geomorphology of the Lower Brazeau River Valley, Foothills of Alberta, Ph.D. Thesis, (in progress), Univ. of Alberta.

Langbein, W.B. et al. 1949. Annual Runoff in the United States. U.S. Geol. Surv. Circ. 52, 14 pp.

Lane, E.W. 1955. The Importance of Fluvial Morphology in Hydraulic Engineering. Proc. Am. Soc. Civ. Eng. v. 81, No. 745, 17 pp.

Leopold, L.B., Wolman, M.A. and Miller, J.P. 1964. Fluvial Processes in Geomorphology. W.H. Freeman and Co., San Francisco, 521 pp.

Luckman, B.H. 1977. Lichenometric Dating of Holocene Moraines at Mount Edith Cavell, Jasper, Alberta. Can. J. Earth Sci. , v. 14(8), pp. 1809-1822.

Luckman, B.H. and Osborn, G.D. 1977. Holocene Glacier Fluctuations in the Middle Canadian Rocky Mountains. Quat. Res., No. 11, pp. 52-77.

McPherson, H. 1963. The Glacial Geomorphology of the Upper Red Deer River Valley, Alberta, M.Sc. Thesis, Univ. of Alberta, 41 pp.

McPherson, H.J. 1968. Historical Development of the Lower Red Deer River Valley, Alberta. Can. Geog., v.12, No. 4, pp. 227-240.

Miall, A.D. 1977. A Review of the Braided River Depositional Environment, Earth Sci. Rev., v. 13, pp. 1-62.

Moss, J.H. 1974. The Relationship of River Terrace Formation to Glaciation in the Shoshone River Basin, Western Wyoming: in D.R. Coates, (ed.), Glacial Geomorphology, State Univ. of New York, Binghampton, pp. 293-314.

Mountjoy, E.W. 1964. Miette Map-Area, Rocky Mountains of Alberta. Geol. Surv. Can. Mem.

Murdoch, J. and Barnes, J.A. 1970. Statistical Tables for Science, Engineering Management and Business Studies. Macmillan Press Ltd., England, 40 pp.

Neill, C.R. 1973. Hydraulic and Morphologic Characteristics of the Athabasca River Near Fort Assiniboine. Res. Council Alberta Open File Rept., 1973-33, 69 pp.

Pheasant, D.A. 1968. Glacial Geomorphology Ya-Ha Tinda Area, Alberta. Ph.D. Thesis, Univ. of Calgary, 106 pp.

Quinn, J.H. 1957. Paired River Terraces and Pleistocene Glaciation. J. Geol., v. 65, pp. 149-166.

Rains, R.B. 1969. Some Aspects of the Fluvial Geomorphology of the Whitemud Basin, Central Alberta. Ph.D. Thesis, Univ. of Alberta, 240 pp.

Roed, M.A. 1968. The Surficial Geology of the Edson Map Area, (83F), Alberta. Ph.D. Thesis, Univ. of Alberta, pp. 200.

Roed, M.A. 1975. Cordilleran and Laurentide Multiple Glaciation West-Central Alberta, Canada. Can. J. Earth Sci., v. 12(9), pp. 1493-1515.

Rutter, N.W. 1972. Geomorphology and Multiple Glaciation in the Area of Banff, Alberta. Geol. Surv. Can. Bull., 206, 54 pp.

Rutter, N.W. 1977. Multiple Glaciation in the Area of Williston Lake, British Columbia. Geol. Surv. Can. Bull., 273, 31 pp.

Schumm, S.A. 1963. The Disparity Between Present Rates of Denudation and Orogeny. U.S. Geol. Surv. Prof. Paper, 454-H, 13 pp.

Schumm, S.A. 1977. The Fluvial System. Wiley Intersci., New York, 338 pp.

Schumm, S.A. and Parker, R.S. 1973. Implications of Complex Response of Drainage Systems for Quaternary Alluvial Stratigraphy. Nature: Physical Science, No. 243, pp. 99-100.

Stalker, A. MacS. 1960. Surficial Geology of the Red Deer-Stettler Map Area. Geol. Surv. Can. Mem., 306.

Stalker, A. MacS. 1963. Quaternary Stratigraphy in Southern Alberta. Geol. Surv. Can. Paper., 62-34, 52 pp.

Stalker, A. MacS. 1968. Geology of the Terraces at Cochrane, Alberta. Can. J. Earth Sci., v. 5, pp. 1455-1466.

Stalker, A. MacS. 1973. Surficial Geology of the Kananaskis Research Forest and Marmot Creek basin region of Alberta. Geol. Surv. Can. Paper, 72-51, 25 pp.

Stalker, A. MacS. 1977. The Probable Extent of Classical Wisconsin Ice in Southern and Central Alberta. Can. J. Earth Sci., v. 14, pp. 2614-2619.

Stene, L. P. 1966. Pleistocene Valley Terraces Along the Athabasca River in the Hinton Area. M.Sc. thesis, Geog. Dept., Univ. of Alberta, 54 pp.

St.-Onge, D.A. 1967. Quaternary Geology and Geomorphology, Whitecourt (83-J) Map-area, Alberta. Geol. Surv. Can. Paper. 67-1, pp. 186-189.

St.-Onge, D.A. 1970. Quaternary Geology and Geomorphology of Tawatinaw Area, Alberta. Geol. Surv. Can. Paper, 70-1 83 pp.

St.-Onge, D.A. 1972. Sequence of Glacial Lakes in North-central Alberta. Geol. Surv. Can. Bull., 213, 16 pp.

Taylor, R.S. 1960. Some Pleistocene Lakes of Northern Alberta and Adjacent Areas. J. Alberta Soc. Pet. Geol., v. 8, No. 6, 167-185.

Thomson, S. and Townsend, D.L. 1978. River Erosion and Bank Stabilization, North Saskatchewan River, Edmonton, Alberta. presented to 31st Can. Geotech. Conf., Winnipeg, 33 pp.

Tokarsky, O. 1977. Hydrogeology of the Whitecourt Area, Alberta. Res. Council Alberta Rept. 76-3, 10 pp.

Visher, A.S. 1965. Use of Vertical Profile in Environmental Reconstruction. Am. Assoc. Pet. Geol. Bull., v. 49, pp. 41-61.

Wagner, W. P. 1966. Correlation of the Rocky Mountain and Laurentide Glacial Chronologies in Southwestern Alberta, Canada. Unpub. Ph.D. thesis, U. of Michigan, 141 pp.

Walcott, R.I. 1970. Isostatic Response to Loading of the Crust in Canada. Can. J. Earth Sci., v. 7, pp. 716-725.

Walker, R. A. 1976. Facies Models 3: Sandy Fluvial Systems. Geosci. Can., v. 3, pp. 101-109.

Weertman, J. 1961. Equilibrium Profile of Ice Caps. J. Glac. v. 3, pp. 953-964.

Westgate, J.A. 1969. The Quaternary Geology of the Edmonton Area, Alberta in Pawluck, S. editor, Pedology and Quaternary Research. Univ. of Alberta Press.

Westgate, J.A., Kalas, L. and Evans, M.E. 1976. Geology of the Edmonton Area, Field Trip Guidebook C-8. Geol. Ass. of Can., Edmonton Geol. Soc. Ann. Meeting, 1976, 49pp.

Wolman, M.A. and Leopold, L.B. 1957. River Flood Plains: Some Observations on Their Formation. U.S. Geol. Surv. Prof. Pap. 282-C.

APPENDIX A

STRATIGRAPHIC SECTIONS

The following descriptions of measured stratigraphic sections are presented starting with the highest and ending with the lowest exposed units. Unit thicknesses are given in meters. Sections are located from the National Topographic Survey (NTS) maps by use of the Military Topographic grid referencing system (MTS).

Section No: 1-1

Location: NTS 83F/5E(837211)

Unit No.	Lithology	Thickness
1	Soil	0.42
2	Gravels Rounded to well rounded gravels, horizontally bedded. Individual beds are indistinct; some beds exhibit a fining upward sequence while others display no visible sorting. Gravels consist primarily of quartzites, limestones, sandstones and conglomerates of Cordilleran origin. Modal pebble size is approximately 2.0 cm with clasts up to 30 cm included. The unit is poorly consolidated, with a medium to coarse sand matrix.	2.4
3	Gravels Subangular to subrounded cross-bedded gravels. Individual beds are short and discontinous, many exhibiting a fining upward sequence. Gravel composition is similar to the upper gravel unit, with a larger pebble mode of approximately 5 cm. Large boulders over 30 cm occur more frequently. The unit is poorly consolidated, with a coarse sand matrix.	2.9
	Total	5.72

Section No. 1A-1

Location: NTS 83F/6W (683253)

Unit No.	Lithology	Thickness
1	Soil	0.31
2	Till Light brown, slightly plastic when moist, moderately stony. Pebble diameter mode is approximately 3.0 cm with clasts up to 50 cm included. Clasts are subangular to subrounded quartzites, limestones, sandstones and conglomerates of Cordilleran origin. The unit has a sandy silt/clay matrix and is moderately consolidated.	2.6
3	Bedrock	-
	Total	2.91

Section No: 1A-2

Location: NTS 83F/6W (677244)

Unit No:	Lithology	Thickness
1	Sands Coarse to medium grained sands, moderately bedded and compacted. Minor amounts of silts and clays are also present.	0.72
2	Gravels Subrounded to rounded gravels displaying no apparent bedding structure. Gravels consist primarily of quartzites, limestones and sandstones of Cordilleran origin. Modal pebble size is approximately 2.3 cm. The unit is	

poorly sorted with a coarse sand matrix. 3.2

Total 3.92

Section No: 1A-4

Location: NTS 83F/6W (704256)

Unit No:	Lithology	Thickness
1	Soil	0.35
2	Till	
	Olive brown, plastic when moist, very stony. Pebble mode is approximately 3.0 cm with clasts up to 40 cm included. Clasts are subangular to subrounded quartzites, limestones and sandstones of Cordilleran origin. The unit has a sandy silt/clay matrix and well consolidated.	3.0
3	Gravels	
	Highly variable subrounded to rounded gravels, fine gravel and sand lenses being interspersed with coarse gravel beds. Gravels consist primarily of quartzites, limestones and sandstones of Cordilleran origin. The unit has a coarse sand matrix and is poorly consolidated. Fluctuating sequences of coarse gravel lenses fining upward to fine gravels and medium to fine grained sand lens involutions are scattered throughout the unit. The pebble diameter mode varies from 2.0 to 5.0 cm depending on the gravel unit, with clasts up to 50 cm included.	15.0
	Total	18.35

Section No: 1A-4-1

Location: NTS 83F/5E (720267)

Unit No:	Lithology	Thickness
1	Colluvium	3.0
2	Till Pale brown, plastic when moist, very stony. Pebble diameter mode is approximately 3.4 cm with clasts up to 35 cm included. Clasts are subangular to subrounded quartzites, limestones and sandstones of Cordilleran origin. The unit has a sandy silt/clay matrix and is moderately consolidated.	4.2
	Total	7.2

Section No: 1A-5

Location: NTS 83F/5E (700251)

Unit No.	Lithology	Thickness
1	Soil	0.32
2	Till Light brown, plastic when moist, very stony. Pebble diameter mode is approximately 3.2 cm with clasts up to 48 cm included. Clasts are subangular to subrounded quartzites, limestones, sandstones and minor amounts of conglomerates all of Cordilleran origin. The unit has a sandy silt/clay matrix and is moderately consolidated.	3.0
	Total	3.32

Section No: 1A-6

Location: NTS 83F/5E (697252)

Unit No:	Lithology	Thickness
1	Soil	1.2
2	Gravels Subrounded to rounded gravels displaying no apparent bedding structure. Gravels consist primarily of quartzites, limestones and sandstones of Cordilleran origin. Modal pebble size is approximately 3.0 cm. The unit is poorly sorted with a coarse to medium sand matrix.	2.0
	Total	3.2

Section No: 1A-7

Location: NTS 83F/5E (695248)

Unit No:	Lithology	Thickness
1	Colluvium	15.0
2	Gravels Subrounded to rounded gravels displaying no apparent bedding. Gravels consist primarily of quartzites and limestones, with some sandstones, of Cordilleran origin. Modal pebble size is approximately 7 cm. The unit is poorly consolidated with a coarse sand matrix.	4.0
	Total	19.0

Section No: 1A-9

Location: NTS 83F/5E (672223)

Unit No:	Lithology	Thickness
1	Soil	0.40
2	Gravel Rounded to well rounded gravels, horizontally bedded. Individual beds are thin and poorly sorted, exhibiting alternating coarse and fine gravel lenses. Gravels consist primarily of quartzites, limestones, sandstones and conglomerates of Cordilleran origin. Modal pebble size varies between 3.3 cm and 4.2 cm. The unit is poorly consolidated with a coarse sandy matrix.	5.4
	Total	5.8

Section No: 2-1

Location: NTS 83F/11W (740286)

Unit No:	Lithology	Thickness
1	Colluvium	4.0
2	Till Light brown, plastic when moist, very stony. Pebble diameter mode is approximately 4.0 cm, with clasts up to 70 cm included. Clasts are subangular to subrounded quartzites, limestones, conglomerates and sandstones of Cordilleran origin. The unit has a sandy loam matrix and is moderately consolidated.	6.2
	Total	10.2

Section No: 2-2

Location: NTS 83F/11W (758306)

Unit No:	Lithology	Thickness
1	Soil	0.72
2	Till Olive brown, plastic when moist, very stony. Pebble diameter mode is approximately 5.0 cm, with clasts up to 63 cm included. Clasts are angular to subrounded quartzites, limestones and conglomerates of Cordilleran origin and local sandstones. The unit has a clay loam matrix and is very well consolidated.	4.5
	Total	5.22

Section NO: 2-4

Location: NTS 83F/11W (829399)

Unit No:	Lithology	Thickness
1	Soil	0.35
2	Colluvium	21.0
3	Till Light brown, slightly plastic when moist, very stony. Pebble diameter mode is approximately 3.0 cm with large clasts of variable sizes included. Clasts are angular to subrounded quartzites and limestones of Cordilleran origin with local sandstones and cherts included. The unit has a sandy loam texture and is not highly consolidated.	14.0
	Total	35.35

Section No: 2-4-1

Location: NTS 83F/11W (833390)

Unit No:	Lithology	Thickness
1	Soil	1.2
2	Gravels Rounded to well rounded gravel displaying no apparent bedding. Gravels consist of quartzites and limestones of Cordilleran origin with some local sandstones included. Modal pebble size is approximately 2.8 cm with clasts up to 13 cm included. The unit is poorly consolidated with a coarse sandy matrix.	21.6
	Total	22.8

Section No: 2-5

Location: NTS 83F/11W (822400)

Unit No:	Lithology	Thickness
1	Soil	0.32
2	Till Light brown, very plastic when moist, moderately stony. Pebble diameter mode is approximately 4.0 cm with variable clast sizes up to 56 cm. Clasts are angular to subangular quartzites and limestones of Cordilleran origin, with some local sandstones and coal fragments. The unit has a fine sand silt/clay matrix and is highly consolidated.	5.4
	Total	5.72

Section No: 2-6

Location: NTS 83F/11W (759335)

Unit No:	Lithology	Thickness
1	Soils	0.5
2	Gravels Angular to rounded gravels displaying no apparent bedding. Gravels consist primarily of quartzites and limestones of Cordilleran origin, with local shale, sandstone and coal fragments throughout the unit. The unit has a high carbonate content. Modal pebble size is highly variable throughout the unit, clasts range from 0.1 cm to 20 cm. The unit is poorly consolidated with a medium to fine sand, silt/clay matrix.	4.2
	Total	4.7

Section No: 3-1

Location: NTS 83F/11E (836402)

Unit No:	Lithology	Thickness
1	Soil	0.45
2	Till Light brown, moderately plastic when moist, very stony. Pebble diameter mode is approximately 3.4 cm with clasts up to 38 cm included. Clasts are subangular to subrounded quartzites, limestones, conglomerates and sandstones of Cordilleran origin. The unit has a sandy silt/clay matrix and is moderately consolidated.	3.6

3	Sands Coarse to medium sands, horizontally bedded lenses fining upward. Individual lenses are thin and discontinuous. Minor coal fragments.	0.45
4	Bedrock	-
	Total	4.5

Section No: 3-1-1

Location: 83F/11E (857398)

Unit No:	Lithology	Thickness
1	Soil	0.15
2	Gravels Subrounded to rounded gravels displaying no apparent bedding structure. Gravels consist primarily of quartzites, limestones and conglomerates of Cordilleran origin, with minor amounts of local sandstones. Modal pebble size is approximately 5.6 cm. The unit is poorly sorted, with a coarse sand matrix.	1.5
	Total	1.65

Section No: 3-3

Location: NTS 83F/11E (846402)

Unit No:	Lithology	Thickness
1	Soil	0.50
2	Till Olive to light brown, slightly plastic when moist, moderately stony. Pebble diameter mode is approximately	

2.5 cm with clasts up to 22 cm included. Clasts are angular to subrounded quartzites and limestones of Cordilleran origin and locally derived sandstones and fractured shale fragments. The unit has a sandy silt/clay loam matrix and is poorly consolidated. 3.0

3	Bedrock	1.8
	Total	5.3

Section No: 3-4

Location: NTS 83F/ 11E (849402)

Unit No:	Lithology	Thickness
1	Soil	0.42
2	Till	

Olive brown, slightly plastic when moist, moderately stony. Pebble diameter mode is approximately 2.3 cm with clasts up to 26 cm included. Clasts are angular to subrounded quartzites and limestones of Cordilleran origin and locally derived sandstone and shale fragments. The unit has a sandy silt/clay matrix and is poorly consolidated. 2.7

3 Sands

Medium to fine sands and silts, horizontally bedded. Individual lenses are very thin and discontinuous. 0.34

4	Bedrock	-
	Total	3.46

Section No: 3-4-1

Location: NTS 83F/11E (858400)

Unit No:	Lithology	Thickness
1	Colluvium	1.2
2	Gravels Subrounded to rounded gravels displaying no apparent bedding. Gravels consist primarily of quartzites and limestones of Cordilleran origin and locally derived sandstones and shales. Modal pebble size is approximately 3.2 cm. The unit is poorly sorted and consolidated with a coarse sand matrix.	2.4
	Total	3.6

Section No: 3-8

Location: NTS 83F/11E (897505)

Unit No:	Lithology	Thickness
1	Soil	0.62
2	Gravels Subrounded to rounded gravels displaying no apparent bedding. Gravels consist primarily of quartzites and limestones of Cordilleran origin, with some locally derived sandstones. Modal clast size is variable, ranging from 3.0 to 8.0 cm. The unit is poorly sorted with a coarse to medium sand matrix.	12.16
3	Bedrock	-
	Total	12.78

Section No: 3-9

Location: NTS 83F/11E (501890)

Unit No:	Lithology	Thickness
1	Soil	0.51
2	Gravels Angular to well rounded gravels, displaying no obvious stratification. Gravels consist primarily of quartzites and limestones of Cordilleran origin, with some locally derived sandstone and shale fragments. Modal pebble size is approximately 5.0 cm, with clasts ranging from 0.8 to 13 cm. The unit is poorly sorted with a medium to fine sand/silt matrix. The unit is moderately consolidated.	2.4
	Total	2.91

Section No: 4-3-1

Location: NTS 83F/14E (908658)

Unit No:	Lithology	Thickness
1	Soil	1.2
2	Gravels Subangular to subrounded gravels displaying no apparent bedding. Gravels consist primarily of quartzites, limestones and conglomerates of Cordilleran origin, with some locally derived sandstones. Modal pebble size is approximately 6.0 cm. The unit is poorly sorted, with a coarse to medium sand matrix.	16.0
3	Bedrock	-
	Total	17.2

Section No: 4-4

Location: NTS 83F/14E (931713)

Unit No:	Lithology	Thickness
1	Colluvium	15.2
2	Till Olive to light brown, slightly plastic when moist and not excessively stony. Pebble diameter mode is approximately 3.4 cm with clast sizes ranging from 2.0 to 30 cm. Clasts are angular to subrounded quartzites, limestones and conglomerates of Cordilleran origin and locally derived sandstones. The unit has a sandy/silt matrix and is not highly consolidated.	3.6
3	Bedrock	-
	Total	18.8

Section No: 6-1

Location: NTS 83K/2 (109843)

Unit No:	Lithology	Thickness
1	Gravels Subrounded to rounded gravels displaying no apparent bedding structure. Gravels consist primarily of quartzites and limestones of Cordilleran origin and locally derived sandstones and shales. Minor traces of igneous and metamorphic clasts of Canadian Shield origin were also identified. The unit is poorly consolidated, with a coarse sand matrix.	1.52
2	Bedrock	-
	Total	1.52

Section No: 6-1-1

Location: NTS 83K/2 (108857)

Unit No:	Lithology	Thickness
1	Soil	0.47
2	Gravels Subrounded to rounded gravels displaying no apparent bedding structure. Gravels consist of quartzites and limestones of Cordilleran origin, local sandstones and shales, with some igneous and metamorphic clasts of Canadian Shield origin. Modal pebble size is approximately 3.6 cm, with clasts ranging from 1.0 to 7.0 cm. The unit is poorly consolidated with a coarse to medium sand matrix.	3.1
	Total	3.57

Section No: 6-2-1

Location: NTS 83K/2

Unit No:	Lithology	Thickness
1	Soil	0.40
2	Gravels Subrounded to rounded gravels displaying no apparent bedding structure. Gravels consist of quartzites and limestones of Cordilleran origin, locally derived sandstones, with some igneous clasts of Canadian Shield origin. Modal pebble size is approximaely 4.5 cm, with clasts ranging from 1.0 to 5.3 cm. The unit is poorly consolidated, with a coarse to medium sand matrix.	5.8
	Total	6.2

Section No: 6-4

Location: NTS 83K/2 (137847)

Unit No:	Lithology	Thickness
1	Soil	0.38
2	Sands Coarse to medium grained sands, with minor gravel inclusions, displaying no apparent bedding. Gravels are confined to the upper 12 cm of the section. Gravel clasts are subrounded to rounded, ranging in size from 2.0 to 12 cm. Gravels consist of quartzites and limestones of Cordilleran origin with some minor igneous fragments of Canadian Shield origin.	3.0
	Total	3.38

Section No: 6-5-2

Location: NTS 83K/2 (114843)

Unit No:	Lithology	Thickness
1	Soil	0.30
2	Till Grey to dark brown, plastic when moist, moderately stony. Pebble diameter mode is variable ranging from 2.0 to 3.0 cm, with boulders up to 2 meters included. Clasts are mainly angular to subrounded Cordilleran quartzites and limestones; however igneous and metamorphic clasts of Canadian Shield origin are also common. The unit has a clay loam matrix and is well consolidated.	4.2
	Total	4.5

Section No: 6-6

Location: 83K/2 (079849)

Unit No:	Lithology	Thickness
1	Soil	0.15
2	Sands Medium to fine sands displaying stratified foreset beds. Minor clay lenses are present but are discontinuous and thin. No gravel clasts were observed. Individual sand lenses fine upwards.	3.6
	Total	3.75

Section No: 6-6-1

Location: NTS 83K/2 (259996)

Unit No:	Lithology	Thickness
1	Soil	0.25
2	Gravels Rounded to well rounded gravels, displaying no apparent bedding structure. Gravels consist primarily of quartzites, limestones and sandstones, with minor amounts of igneous and metamorphic clasts of Canadian Shield origin included. Modal pebble size is approximately 6.4 cm, with clasts ranging from 0.8 to 12 cm included. The unit shows little evidence of sorting with a coarse sand matrix.	2.4
3	Bedrock	-
	Total	2.65

Section No: 6-7

Location: 83K/2 (094845)

Unit No:	Lithology	Thickness
1	Soil	0.30
2	Gravels Subrounded to rounded gravels with no apparent bedding structure. Gravels consist primarily of quartzites and limestones with some igneous and metamorphic clasts of Canadian Shield origin. Modal pebble size is approximately 6.5 cm, with clasts ranging between 1.5 to 12 cm. Occasional medium sand lens involutions were evident; however individual lenses are thin and discontinuous throughout the unit. The unit is poorly consolidated with a coarse to medium sand matrix.	9.12
	Total	9.42

Section No: 6-11

Location: NTS 83K/2 (260007)

Unit No:	Lithology	Thickness
1	Soil	0.60
2	Gravels Rounded to well rounded gravels, displaying no apparent bedding structure. Gravels consist primarily of quartzites, limestones and sandstones, with very few igneous or metamorphic clasts of Canadian Shield origin. Modal clast size is variable, ranging from 2.0 to 7 cm. The unit is poorly consolidated, with a coarse to medium sand matrix.	1.8
3	Bedrock	-
	Total	2.40

Section No: 6-12

Location: NTS 83K/2 (254024)

Unit No:	Lithology	Thickness
1	Sands Medium to fine grained sands displaying no apparent bedding structure. No gravels are present within the unit, however minor traces of coal fragments are visible.	Variable
2	Till Light to dark grey brown, plastic when moist, very stony. Pebble mode is variable-ranging from 2.0 to 3.0 cm with boulders up to 70 cm included. Clasts are angular to subrounded quartzites, limestones and sandstones with a high percentage of igneous and metamorphic clasts of Canadian Shield origin included. The unit has a clay loam matrix and is well consolidated.	2.4
3	Gravels Subrounded to rounded gravels displaying no apparent bedding structure. Gravels consist almost exclusively of quartzites and limestones with only a few igneous and metamorphic clasts of Canadian Shield origin included. Modal pebble size is approximately 10 cm, but sizes vary from 1.5 to 25 cm. The unit is very poorly consolidated with a coarse sand matrix. The contact zone with the overlying tills is very distinct.	3.1
4	Bedrock	-
	Total	5.5

Section No: 6-20

Location: NTS 83K/2 (268002)

Unit No:	Lithology	Thickness
1	Soil	0.24
2	Gravels	

Subrounded to rounded gravels, with no apparent bedding structure. Gravels consist primarily of quartzites, limestones and sandstones, with only a few igneous and metamorphic clasts of Canadian Shield origin included. Modal pebble size is approximately 3.9 cm, with clasts ranging from 1.5 to 12 cm. The unit is poorly sorted, with a medium to fine sand/silt matrix. 1.97

Total 2.21

Section No: 6-30

Location: NTS 83K/7 (257122)

Unit No:	Lithology	Thickness
1	Soil	0.30
2	Till	

Dark grey brown, plastic when moist, moderately stony. Pebble diameter mode is variable, clasts vary from 1.0 to 30 cm, with boulders up to 85 cm included. Clasts are angular to subrounded quartzites, limestones and igneous and metamorphic clasts of Canadian Shield origin. Local sandstones are also present. The unit is very well consolidated, exhibiting a columnar structure, with a clay-rich matrix. 4.56

Total 4.86

Section No: 6-33

Location: NTS 83K/2 (084871)

Unit No:	Lithology	Thickness
1	Soil	0.30
2	Sands Orange/brown coarse sands, horizontally bedded. No gravels are present. A sequence of calcium carbonate bands leached from overlaying soils run vertically through the unit.	1.2
3	Gravels Thin lenses of subrounded to rounded gravels. Gravels consist of quartzites and limestones. Clast size varies from 0.5 to 2.0 cm. The unit is poorly sorted and discontinuous.	0.15
4	Sands Medium to fine sands and silts light grey/brown, horizontally bedded. Individual lenses display no apparent stratification. Several clay lenses are present throughout the unit. No gravels are present	2.4
	Total	4.2

Section No: 6-35

Location: NTS 83K/2 (173837)

Unit No:	Lithology	Thickness
1	Soil	0.30
2	Till Light grey brown, slightly plastic when moist, moderately to very stony. Clast size is variable, ranging from 2.0 to 43 cm. Clasts are angular to	

subrounded quartzites and limestones with a few igneous and metamorphic clasts of Canadian Shield origin included. The unit has a sandy silt/clay matrix and is moderately consolidated. 3.6

Total 3.9

Section No: 7-2

Location: NTS 83K/1 (340052)

Unit No:	Lithology	Thickness
1	Soil	0.37
2	Gravels	

Subrounded to rounded gravels, displaying no apparent bedding structure. Gravels consist of sandstones, quartzites and limestones with very few igneous or metamorphic clasts and Canadian Shield origin. Modal pebble size is approximately 4.3 cm with clasts ranging from 1.0 to 6.0 cm. The unit is poorly sorted, with a coarse to medium sand matrix. 1.8

Total 2.2

Section No: 7-2-1

Location: NTS 83K/1 (498095)

Unit No:	Lithology	Thickness
1	Soil	0.29
2	Gravels	

Subrounded to rounded gravels displaying no apparent bedding structure. Gravels consist of quartzites and limestone with an increased percentage of sandstones. Few igneous or metamorphic clasts are evident. Modal

pebble size is approximately 4.1 cm, with clasts ranging from 1.5 to 12 cm. The unit is poorly consolidated with a medium sand matrix. 4.5

Total 4.79

Section No: 7-3

Location: NTS 83K/1 (330040)

Unit No:	Lithology	Thickness
1	Gravels	
	Subrounded to rounded gravels displaying no apparent bedding. Gravels consist of sandstones, quartzites and limestones, with a few igneous and metamorphic clasts. Modal clast size is approximately 3.3 cm with pebbles ranging from 1.0 to 5.5 cm. The unit is poorly sorted with a medium sand matrix.	2.1
	Total	2.1

Section No: 7-5

Location: NTS 83K/1 (604064)

Unit No:	Lithology	Thickness
1	Soil	0.52
2	Gravels	
	Rounded to well rounded gravels displaying no pronounced bedding structure. Gravels consist primarily of sandstones, quartzits and limestones. No igneous or metamorphic clasts were evident. Modal pebble size is approximately 4.2 cm, with clasts ranging from 1.0 to 6.0 cm. The unit is poorly consolidated, with a medium sand matrix.	6.9
	Total	7.42

Section No: 7-31

Location: NTS 83K/1 (513078)

Unit No:	Lithology	Thickness
1	Colluvium	4.2
2	Till Dark grey brown, plastic when moist, moderately stony. Pebble mode is variable ranging between 1.0 and 6.0 cm with clasts up to 70 cm included. Clasts are angular to subrounded igneous and metamorphic clasts of Canadian Shield origin and quartzites and limestones, with red ironstone clasts also included. The unit has a silt/clay matrix, is well consolidated, displaying a dense columnar structure.	7.6
3	Bedrock	-
	Total	11.8

Section No: 8-1

Location: NTS 83J/4W (664023)

Unit No:	Lithology	Thickness
1	Sands Medium to fine sands, light yellow/brown, horizontally bedded. The unit is moderately consolidated. No gravels are present, minor coal fragments are evident.	Variable
2	Lacustrine Clays Fine sand, silt/clays, fining upward to silts and silt/clays. The unit is horizontally bedded, well bedded, plastic when moist. Horizontal beds are approximately 1.0 cm extending through the unit. A few isolated occurrences of float stones are evident. These are	

	exclusively quartzites.	2.3
		-
	Total	> 2.3

Section No: 8-2

Location: NTS 83J/4W (657024)

Unit No:	Lithology	Thickness
1	Soil	0.36
2	Gravels Subangular to rounded gravels displaying no apparent bedding structure. Gravels consist primarily of sandstones, quartzites and limestones. Modal clast size is approximately 3.3 cm with clasts ranging from 1.0 to 30 cm. The abundant sandstone clasts are angular in shape, compared to the rounded quartzites. Many coal fragments are also evident. The unit is poorly consolidated with a coarse to medium sand matrix.	2.2
	Total	2.56

Section No: 9-1

Location: NTS 83J/4E (838994)

Unit No:	Lithology	Thickness
1	Soils	0.41
2	Lacustrine clays Medium sands fining upward to silts and silt/clays, horizontally bedded. The unit is well compacted and plastic when moist. Medium grained sand lenses predominate at the base of the unit, and fine upward to alternating light and dark silt and clay lenses. Lenses are approximately 1.0 to 1.5 cm in thickness.	4.5

3 Gravels

Subrounded to rounded gravels displaying no apparent bedding structure. Gravels are almost exclusively quartzites, limestones and sandstones with a high coal fraction and minor amounts of igneous and metamorphic clasts. Modal pebble size is approximately 3.2 cm, with clasts ranging between 1.5 and 30 cm. Gravels fine upward to the contact with the overlying lacustrine sediments, this contact zone is well defined. The unit is poorly consolidated with a coarse sand matrix.

2.4

Total 7.31

Section No: 9-2

Location: NTS 835/4E (835007)

Unit No: Lithology Thickness

1 Sands

Medium to fine grained sands, light brown, horizontally bedded. No gravels are present but scattered coal fragments do occur. The unit is moderately consolidated.

0.76

2 Gravels

Subrounded to rounded gravels fining upward to overlying sands. Gravels are primarily quartzites and sandstones with a few metamorphic clasts included. Modal pebble size is approximately 4.1 cm, with clasts ranging from 1.0 to 8.0 cm. The unit is poorly consolidated, with a medium to fine sand matrix.

1.2

Total 1.96

APPENDIX B

Field and Data Analysis Procedures Used to Obtain the Predicted Median and Mean Sieve Diameters, d_{sp50} and $\bar{d}_{sp}$

1. <u>Site Selection</u>

   (a) The clasts associated with the terrace gravel deposits would be approximately ellipsoidal in shape.

   (b) Because numerical method predictions are for isotropic materials terrace gravel deposits which appear quite isotropic are preferable.

   (c) Grain-size analysis should be performed in this manner for discrete beds.

2. <u>Terrace Gravel Grid Placement and Photography</u>

   (a) The section face should be modified to give a reasonably even surface in order to decrease photographic scale distortion.

   (b) For clarity the gravel surface may be sprayed white or colour film may be used.

   (c) If the terrace gravels are extremely coarse grained a frame with attached rulers, but no grid, may be used. The grid pattern may be superimposed on the print later.

   (d) Each grid should be labelled in some way.

3. <u>Grain Selection on Prints</u>

   (a) Fifty to 100 grains should be selected. This may require more than one grid photograph.

(b) Selected grains $\geq$ 8 mm. found under the grid intersection points are measured. If a clast lies under two intersection points it must be counted twice (Kellerhals and Bray, 1971).

4. Grain Measurement from the Prints

(a) Measure the apparent major axis a_t, and apparent minor axis b_t, of each selected grain with calipers.

(b) Determine the scale factor from the grid rulers and apply it to the apparent axes values.

5. Primary Data Analysis

(a) The frequency of the apparent major and minor axes data respectively is determined by using 0.25ϕ or 0.50ϕ class intervals.

(b) The cumulative size-frequency distribution is plotted on arithmetic probability paper.

(c) The Ø16, Ø50 and Ø84 values of the distributions in (b) are noted. The Ø50 values of the apparent major and minor distributions are the values of a_{t50} and b_{t50}, respectively. The Folk and Ward (1957) mean $\frac{\text{Ø}16 + \text{Ø}50 + \text{Ø}84}{3}$ of the apparent major minor distributions yields the mean values $\bar{a}_t$ and b_t, respectively.

6. <u>The Numerical Method Median and Sieve Values</u>

The following presents the method of computing the predicted median sieve diameter, d_{sp5o}.

(a) Convert a_{t50} and b_{t50} values from ϕ units to mm.

(b) Calculate C_{p5o}:

Using $k_2 = \frac{b_{t50}}{a_{t50}}$ in Figure 4-3, $\frac{(\bar{b}-c)100}{c} = y$,

$C_{p5o} = (1.0 - y)(b_{t50})$

$\bar{C}_p$ is calculated in the same way except $\bar{b}_t$ and $\bar{a}_t$ are used.

(c) Calculate d_{sp50}:

$$d_{sp50} = \frac{C_{p50}}{2k_2}\left[\frac{2(1 + k^2)}{2}\right]^{1/2}$$

where $k_2 = \frac{b_{t50}}{a_{t50}}$

$\bar{d}_{sp}$ is calculated in the same way except $\bar{b}_t$, $\bar{a}_t$ and $\bar{C}_p$ are used.

Source:
Bramm (1977)

APPENDIX C

Examples of Predicted Median and Mean Grain Size Calculations for Ten Samples Selected at Random From a Total of Thirty-two Analysed

Sample No.	Axes	Median Sieve Diameter (mm) ϕ values	mm values	$C_{p50}=(1.0-y)(b_{t50})$	$d_{sp50}=\frac{C_{p50}}{2k_2}[2(1+k_2^2)]^{1/2}$	Axes	Mean Sieve Diameter (mm) ϕ Values	mm values	$c_{p50}=(1.0-y)(\bar{b}_t)$	$\bar{d}_{sp}=\frac{\bar{c}_p}{2k_2}[2(1+k_2^2)]^{1/2}$
1-1 Lower	a_{t50}	-5.90	59.71	41.66	55.33	$\bar{a}_t$	-6.0	63.97	41.36	54.85
	b_{t50}	-5.23	37.53			$\bar{b}_t$	-5.34	40.55		
1-1 Upper	a_{t50}	-4.46	22.01	14.85	20.14	$\bar{a}_t$	-4.54	23.28	14.93	20.27
	b_{t50}	-3.75	13.45			$\bar{b}_t$	-3.83	14.22		
1A-3	a_{t50}	-5.26	38.23	28.24	36.25	$\bar{a}_t$	-5.28	38.82	32.61	40.60
	b_{t50}	-4.66	25.28			$\bar{b}_t$	-4.74	26.73		
1A-9	a_{t50}	-5.28	38.85	26.87	35.89	$\bar{a}_t$	-5.29	39.08	25.26	33.49
	b_{t50}	-4.60	24.25			$\bar{b}_t$	-4.63	24.77		
3-4	a_{t50}	-4.72	26.35	19.95	24.24	$\bar{a}_t$	-4.82	28.25	20.06	26.03
	b_{t50}	-4.15	17.75			$\bar{b}_t$	-4.19	18.24		
6-5-1	a_{t50}	-5.45	43.71	30.99	40.79	$\bar{a}_t$	-5.63	49.55	35.43	45.98
	b_{t50}	-4.80	27.86			$\bar{b}_t$	-5.01	32.21		
6-36	a_{t50}	-5.03	32.67	25.56	31.73	$\bar{a}_t$	-5.19	36.48	29.32	37.25
	b_{t50}	-4.50	22.63			$\bar{b}_t$	-4.61	24.43		
7-2-1	a_{t50}	-5.45	43.71	32.01	41.31	$\bar{a}_t$	-5.43	43.15	33.58	43.11
	b_{t50}	-4.84	28.64			$\bar{b}_t$	-4.83	28.46		
7-5	a_{t50}	-5.48	44.63	33.57	42.68	$\bar{a}_t$	-5.46	44.06	40.46	49.37
	b_{t50}	-4.90	29.86			$\bar{b}_t$	-4.96	31.12		
9-1	a_{t50}	-5.15	35.51	24.55	32.79	$\bar{a}_t$	-5.16	35.73	23.10	31.64
	b_{t50}	-4.47	21.61	24.55		$\bar{b}_t$	-4.50	22.65		

APPENDIX D

Formulas Applied in Homogeneity of Variance and Difference of Means Tests

APPENDIX D

Formulas Applied in Homogeneity of Variance and Difference of Means Tests

Homogeneity of Variance (F-test)

$$F = \frac{S_x^2}{S_y^2}$$ where S_x^2 and S_y^2 = sample variances

(Note: S_x^2 is always taken as the largest variance).

and $F_c = F_{\alpha}, \nu_1(n-1), \nu_2(n-1)$ where α = level of significance

ν_1 = degrees of freedom in the numerator

ν_2 = degrees of freedom in the denominator

n = sample size

Difference of means (t-test)

$$t = \frac{\bar{x}_1 - \bar{x}_2}{\sqrt{\frac{(n_1-1)S_x^2 + (n_2-1)S_y^2}{n_1 + n_2 - 2}\left(\frac{1}{n_1} + \frac{1}{n_2}\right)}}$$

where $\bar{x}_1$ and $\bar{x}_2$ = sample means

(Note: All other symbols are as previously defined)

and $t_c = t_{\gamma}, \nu$ where $t_{\gamma} = n_1+n_2 - 2$

See: Cole and King (1968); Murdoch and Barnes (1970)